

Vol. 25, No. 1, January-February, 1990

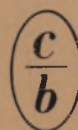
September, 1990

LITHOLOGY AND MINERAL RESOURCES

ЛИТОЛОГИЯ И ПОЛЕЗНЫЕ ИСКОПАЕМЫЕ

(LITOLOGIYA I POLEZNYE ISKOPAEMYE)

TRANSLATED FROM RUSSIAN



CONSULTANTS BUREAU, NEW YORK

LITHOLOGY AND MINERAL RESOURCES

Lithology and Mineral Resources is abstracted or indexed in *Chemical Abstracts*, *Engineering Index*, and *Metal Abstracts Index*.

Mailed in the USA by Publications Expediting, Inc., 200 Meacham Avenue, Elmont, NY 11003.

POSTMASTER. Send address changes to *Lithology and Mineral Resources*, Plenum Publishing Corporation, 233 Spring Street, New York, NY 10013.

Lithology and Mineral Resources is a translation of *Litologiya i Poleznye Iskopaemye*, a publication of the Academy of Sciences of the USSR.

An agreement with the Copyright Agency of the USSR (VAAP) makes available both advance copies of the Russian journal and original glossy photographs and artwork. This serves to decrease the necessary time lag between publication of the original and publication of the translation and helps to improve the quality of the latter. The translation began with the first issue of the Russian journal.

Editorial Board of *Litologiya i Poleznye Iskopaemye*:

Editor: V. N. Kholodov

Associate Editors: B. M. Mikhailov and P. P. Timofeev

Secretary: G. Yu. Butuzova

A. N. Dmitrievskii	A. A. Migdisov	E. F. Shnyukov
A. V. Il'in	I. O. Murdmaa	S. A. Sidorenko
V. I. Kononov	V. I. Sedletskii	I. I. Volkov
A. I. Konyukhov	E. M. Shmarinovich	O. V. Yapaskurt

Copyright © 1990, Plenum Publishing Corporation. *Lithology and Mineral Resources* participates in the Copyright Clearance Center (CCC) Transactional Reporting Service. The appearance of a code line at the bottom of the first page of an article in this journal indicates the copyright owner's consent that copies of the article may be made for personal or internal use. However, this consent is given on the condition that the copier pay the flat fee of \$12.50 per article (no additional per page fees) directly to the Copyright Clearance Center, Inc., 27 Congress Street, Salem, Massachusetts 01970, for all copying not explicitly permitted by Sections 107 or 108 of the U.S. Copyright Law. The CCC is a nonprofit clearinghouse for the payment of photocopying fees by libraries and other users registered with the CCC. Therefore, this consent does not extend to other kinds of copying, such as copying for general distribution, for advertising or promotional purposes, for creating new collective works, or for resale, nor to the reprinting of figures, tables, and text excerpts. 0024-4902/90/\$12.50

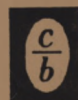
Consultants Bureau journals appear about six months after the publication of the original Russian issue. For bibliographic accuracy, the English issue published by Consultants Bureau carries the same number and date as the original Russian from which it was translated. For example, a Russian issue published in December will appear in a consultants Bureau English translation about the following June, but the translation issue will carry the December date. When ordering any volume or particular issue of a Consultants Bureau journal, please specify the date and, where applicable, the volume and issue numbers of the original Russian. The material you will receive will be a translation of that Russian volume or issue.

Subscription (6 Issues):

Vol. 24: \$855 (domestic); \$990 (foreign)
Vol. 25: \$915 (domestic); \$1060 (foreign)

Single issue: \$150
Single Article: \$12.50

CONSULTANTS BUREAU, NEW YORK AND LONDON



233 Spring Street
New York, New York 10013

Published bimonthly, consisting of Volume 24 issues 3 through 6 and Volume 25 issues 1 and 2. Subscription price for Volume 24 is \$855, and for Volume 25 is \$915. Second-class postage paid at Jamaica, New York 11431.

LITHOLOGY AND MINERAL RESOURCES

A translation of *Litologiya i Poleznye Iskopaemye*

September, 1990

Volume 25, Number 1

January-February, 1990

CONTENTS

Engl./Russ.

Cryolithogenesis and Its Characteristics - I. D. Danilov	1	3
Early Archean History of the Earth and Aquatic Sediment Accumulation		
- A. N. Kazakov	9	13
Isotopic Composition of Sulfidic Sulfur of the Zazhagin Shungite Deposit		
(South Karelia) - G. V. Shatskii	15	20
Ancient Arid Eluvium of Tianshan - V. S. Erofeev and Yu. G. Tsekhovskii	22	29
Landslides and the Formation of Homogeneous Horizons in Submerged Debris Cones		
(Liassic, Northeastern Caucasus) - Yu. O. Gavrilov	38	49
Anomalies in Lower Maikopian Bituminous Clays of the Central Precaucasus		
- A. V. Bochkarev and V. N. Evik	46	59
Paleogeography of the Colchis Lowland Peat Area during the Holocene		
- L. I. Bogolyubova	54	69
Basic Patterns of Terrigenous Sedimentation at the Mouths of Rivers in Arid and		
Semiarid Zones - Yu. P. Khrustalev	71	91
γ -Thermoluminescence Method for Studying Carbonate Deposits - V. D. Il'in, L. N. Klenina,		
and V. N. Rumakin	83	105

*The Russian press date (podpisano k pechati) of this issue was 1/2/90.
Publication therefore did not occur prior to this date, but must be assumed
to have taken place reasonably soon thereafter.*

Sedimentary rock formation in cryolithologic zones during hypergenesis, sedimentogenesis, and diagenesis is described. A special zonal type of lithogenesis is identified - cryolithogenesis. The cryogenetic process (freezing and underground ice formation) is a component of this process. It appears at different stages and is affected by the facies-genetic sediment type.

To a large extent, our planet is made up of ice. According to Shumskii and Krenke [11], ice sheets and glaciers amount to some 16 million km², i.e., 11% of land. Permafrost rocks with ground ice (outside glaciers) make up approximately an equal amount, and sea ice covers an average of 26 million km², i.e., 7% of the ocean area, without aquatoria containing separate icebergs. In the recent geologic past during the Late Pleistocene, just 18-20,000 years ago, the spread of ice on the earth almost doubled. Natural ice and the specific sedimentation settings associated with them covered huge expanses on the surface of the planet during the Paleozoic, the Late and Early Proterozoic, and possibly also during the Late Archean.

Recognizing the special role of ice in the sedimentary process, Strakhov described a glacial type of lithogenesis among the main climatically determined types, along with humid and arid. By definition, it is associated with areas of widespread ice sheets that existed for a geologically long time rather than with any particular climatic zone. Shilo ([10], etc.) described the concept of periglacial lithogenesis in addition to the ice (glacial) type. While the latter is typical for regions with a negative thermal balance and a positive balance of solid meteoric sediments (snow), the former is common where two conditions are observed: a deficit of heat and a deficit of solid meteoric precipitation. Glacial lithogenesis exists in regions with a cold and moist climate; periglacial lithogenesis exists in regions with cold arid climates. The term "polar lithogenesis" also is used by lithologists [1, 6]. It is a zonal description that refers to polar and subpolar regions with a negative thermal balance and active participation of ice in the sedimentary process. Polar lithogenesis exists not only on continents but also in sea basins - the terminal runoff receptacles of sedimentary matter from adjacent land with land glaciation (the Antarctic) or both land and underground glaciation (northern Eurasia and North America).

The term "cryolithogenesis" has been used increasingly in the past few years. There is no consensus as to its definition. Some authors consider cryolithogenesis to be merely the process of freezing and thawing of rocks associated with the formation and disappearance of ground ice [8]. An alternative interpretation defines cryolithogenesis as the formation of frozen, mainly Quaternary sediments [5] or as a peculiar sediment process [2, 4]. However, most authors take the term in the sense of rock freezing and ground ice formation, both perennial and stable (permanent or multiannual) and seasonal or cyclic. This interpretation does not include transport of sedimentary matter or its accumulation and diagenetic alteration in terminal discharge basins.

Freezing and ground ice formation can be viewed as an element of cryolithogenesis which can be called "cryogenesis." Cryolithogenesis is a much broader concept. Its manifestations cover both continental and marine blocks of the earth in polar and subpolar altitudes and portions of dry land with harsh climates in lower latitudes as well. In this definition, cryolithogenesis comprises the following types of lithogenesis as described previously in association with the various forms of existence of ice on the earth:

- 1) above-ground - glacial lithogenesis, widespread in areas of ice sheets that existed for a geologically long time [9];
- 2) underground - periglacial lithogenesis, common in areas of permafrost rocks on land [10];

Moscow State University. Translated from *Litologiya i Poleznye Iskopaemye*, No. 1, pp. 3-12, 1990. Original article submitted January 19, 1989.

3) marine - polar lithogenesis, in cold water icy seas and oceans with mostly negative bottom temperatures [1, 6].

This definition of cryolithogenesis is based on the theoretical concepts of the cryosphere developed by Doborovol'skii and Vernadsky. The cryosphere is the shell of the planet, which comprises components of the atmosphere, hydrosphere, and lithosphere with negative temperatures. Cryolithogenesis evolves in the latter two. Traditionally, perennially frozen beds of the lithosphere and ground ice have been studied by geocryology, above-ground glaciers, and by glaciology. Yet, it is obvious that both objects belong within the purview of lithology. The lithologic significance of glaciers is well known. The lithologic effect of underground ice formation is tremendous, although it is yet to be fully appreciated by lithologists. Sumgin, the founder of modern geocryology, and later Doborovol'skii and Vernadsky, insisted on the importance of studying the entire cryosphere of the earth. Elaborating on these principles, Koloskov included in the cryosphere beds of permafrost rocks with ground ice, glacial ice, water layers in seas with negative temperatures, and floating ice. This prepared the way for the modern concept of the earth's cryolithozone as the domain of cryogenic lithogenesis, or cryolithogenesis [3].

Cryolithogenesis can be viewed as a special climatically determined type of subtype of sedimentary rock formation if it is characterized by zonality and specificity. Arid lithogenesis types were defined on the basis of these principles. For glacial lithogenesis Strakhov [9] made an exception, because ice sheets in the Northern Hemisphere do not constitute a contiguous zone but form separate foci or regions: Greenland, the Canadian Arctic archipelago, and the archipelagos of the Spitsbergen, Franz Joseph Land, Northern Land, and Novaya Zemlya. In our definition of cryogenesis, the areas where it is found correspond to all three forms of glaciation on the earth: above-ground, underground, and marine (i.e., the cryolithozone). The zonality of cryolithogenesis in this case is obvious, but its specificity requires special discussion. It is determined by two main factors: low (negative average annual) temperatures and participation of ice in various phases of the sedimentary process (mobilization of matter during weathering, its transport and accumulation, and conversion to frozen sedimentary rock).

Weathering (hypergenesis) of rocks in permafrost areas is largely contributed to by the cryogenic factor and is even described by the terms cryogenic weathering and cryogenic eluvium. As monolithic hard rocks become fractured, the wedging effect of water that freezes to ice becomes significant. Large piles of clastic eluvium - boulder and scree-boulder - material are formed. Fracturing is accompanied by film cryohydration, and the physicochemical and biochemical processes producing fine soil: all kinds of combinations of psammitic, siltic, and pelitic particles. In the eluvium, the material is sorted cryogenically. Larger fragments bulge upward, forming rock rings and polygons. The base of the eluvium becomes enriched in fine ground. The final result of cryogenic weathering is granulometrically uniform loess-like loam in subsurface occurrence, sometimes described as cryopelite [8]. On planes they form cloak-like covers up to 3-5 m thick, sometimes enveloping watersheds, slopes, and terraces. The base of loess-like covering loams is uneven and forms wedge-shaped structures closely correlated with lows in so-called permafrost polygonal-block relief, consisting of alternating convex quadrangular blocks with smoothed edges, separated by interblock lows. The breaking of rocks in the cryolithozone is a peculiar process, producing a special kind of eluvium: a cryogenic crust of weathering.

The transport of sedimentary matter in the cryolithozone is also specific and observed both in subaerial and subaquatic sedimentation settings. Slope processes are well known in cryolithozones: solifluxion, creep, stone stream formation, etc. These processes are intensive because of the shallow occurrence of the roof of permafrost rocks under seasonally thawing subsurface layers. The cryogenic specific transport of sedimentary matter by water flows and in terminal discharge basins are less known. Small water streams virtually do not sort clastic matter they receive from slopes. Sometimes it is sorted in valleys even worse than on slopes [10]. This phenomenon has been described as "valley" of "cryogenic alluvium." The specific transport of sedimentary matter by large rivers is affected by surface ice, especially noticeable in Siberian rivers flowing from south to north. Large backwaters are formed during high-water seasons, resulting in local tides up to 18-20 m in the lower reaches of the Yenisei River.

River ice carries large amount of fine grounds and rock fragments up to 3 m across. In terminal discharge basins (seas and lakes) floating ice is an even stronger factor of transport of clastic matter than in rivers. Given the occurrence area of floating ice mentioned

earlier, the importance of this transport should be immense. The amount of clastic matter carried by ice has been estimated by investigators for polar sea sediments as 1-2 tons/m³. Gravel, pebble, and even small boulders have been found in seafloor sediments in the central area of the Arctic Basin. They are brought there with fast ice, which accounts here for some 20% of surface ice sheets. According to Leont'ev's estimates [7], floating ice annually supplies to the Arctic Ocean 180 million tons of sedimentary matter, which is almost equal to the amount discharged by rivers (200 million tons). The bulk of the material (about a billion tons) is supplied by thermal abrasion. There is a clear difference in supply sources of the Arctic and the world ocean. The latter receives most of the discharge (some 80%) from rivers. The specifics of sediment transport by glaciers are well known. In the cryolithozone, low temperatures and an abundance of organics (because of low intensity of decay) in surface and sub-surface raises the geochemical mobility of elements such as Fe, Mn, and Al. These waters also have higher concentrations of dissolved CO₂ and are aggressive [1, 4]. Transport of solid and dissolved sedimentary matter in the cryolithozone is highly peculiar both on land (in subaerial and subaquatic settings) and in the sea. The reasons for these specific features are low temperatures and ice in its various forms: above-ground, underground, and marine.

Accumulation of sedimentary matter is affected by the transport mechanism and thermal conditions in the environment. In coldwater polar seas bottom sediments are almost entirely terrigenous; there are no biogenic and chemogenic varieties. The ice factor forms a peculiar type of finely dispersed water sediments: poorly sorted clay and loam with large detrital inclusions, and, in particular, a typical diamicton of ice-marine (cryomarine) origin (Fig. 1a). The main laws of sedimentology are violated in Arctic polar basins. One fundamental principle states that the sea is the terminal discharge basin where differentiation of terrigenous matter evolves to its fullest extent [9]. By contrast, sediments of polar basins are reminiscent of unsorted glacial accumulations [1, 2]. Glacial and marine sediments, which are normally diametrically opposed in terms of degree of sorting, are similar in the cryolithozone. In both situations, their composition is controlled by the cryogenic sedimentation factor, i.e., ice. Generally, one can speak of cryogenic sedimentogenesis in marine and continental facies.

On continents, surface ice of water basins also controls sedimentation. Rock fragments carried by river ice enriched finely dispersed flood valley sediments, such as silty sand-loam (see Fig. 1b). This lithologic variety of alluvial sediments is characteristic of the cryolithozone and not found in other lithogenesis zones. A series of glacial-alluvial facies can be defined amid flood valley sediments: boulders, blind areas on river waterfronts, shore bars formed of boulders and pebbles, and river flood valleys composed of boulder sand-loam. The specifics of sedimentation in the cryolithozone are present in this process even before freezing and underground ice formation (cryogenesis), which of course make it so peculiar. However, even without this latter process certain genetic sediment types have cryolithogenic individuality, which makes them quite different from their counterpart sediments in other lithogenesis zones. A genetic interpretation based on lithogenesis patterns described for a humid (warm and moderate) climate introduces basic errors when beds of ice-marine clays and loams hundreds of meters thick are interpreted as continental land accumulations (moraines). Alluvial-flood valley boulder finely dispersed sediments of high and sometimes low river terraces of recent and especially ancient cryolithozones are also often classified as moraines. The failure to consider the cryogenic property of alluvial sedimentogenesis and marine sedimentation is responsible for many incorrect paleogeographic reconstructions. For example, finds of boulders in flood valley facies of river alluvia of the Pechora and West Siberian planes has given rise to theories that the latest glaciation in these regions was of a cover type and buried erosional forms at the level of second and even first flood valley terraces.

The cryogenic factor in the form of floating surface ice does not always result in accumulation of unsorted or poorly sorted sediments with large clastic inclusions. Banded clays and silts well differentiated granulometrically into layers are also common sediment types in basins of the cryolithozone. They are formed in regions with harsh continental climates where solid matter is discharged into closed or semiclosed basins with a pronounced differentiation among seasons. A stable long-term ice sheet that for most of the year protects seafloor sediments even at a relatively shallow depth from wave action is an additional factor. Banded clays and silts are formed not only in undrained or slow-flow subglacial basins but also in relatively deep lakes of the cryolithozone not associated with glaciers and in lagoons, estuaries, and seafloor lows. The latter is indicated by the fauna of marine mollusks, foraminifera, and ostracods.

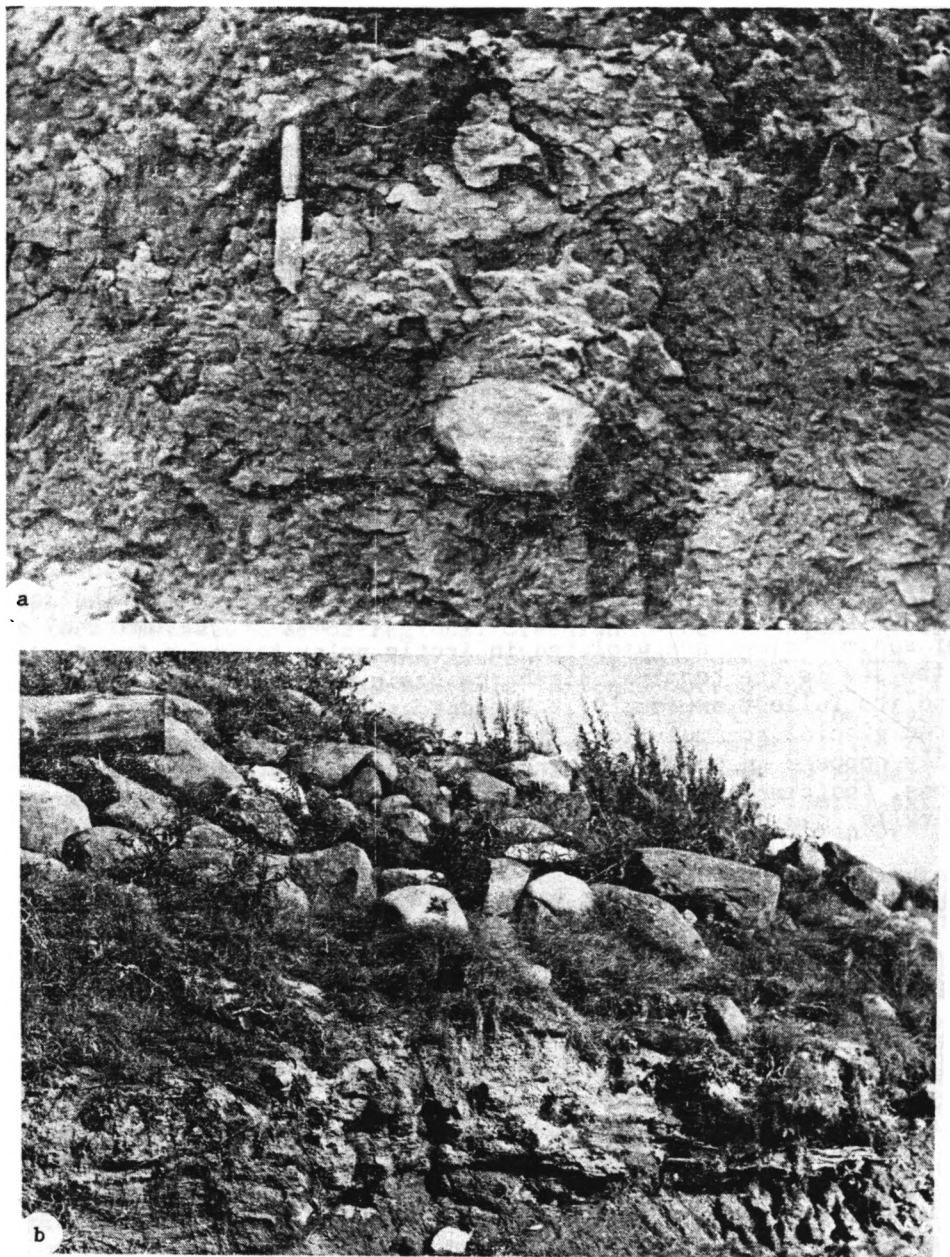


Fig. 1. Poorly sorted boulder deposits: a) glacial-marine loams of the "diamicton" type on the Pechora seashore; b) flood valley peat bog sand loams of a low annually flooded terrace of the Yenisei River (bottom) with piles of rock fragments and flotsam.

Before freezing, basin sediments pass through the stage of ordinary diagenesis, which in terminal discharge basins of the cryolithozone has all the elements of that process: dehydration, compaction, and geochemical and mineralogic transformation. However, unlike the diagenetic processes in the subzone of moderate humid climate, the chemical/mineralogic transformations in the cryolithozone are slowed down by low temperatures. In the finely dispersed varieties of basal sediments that normally contain 0.5-1.0% of reactive organic matter, diagenetic minerals of the sulfide series are present as a finely dispersed impurity: hydrotroilite, pyrite, and iron phosphate (vivianite) [1,2]. Sometimes they form nodular contractions (Fig. 2a). In a sharply continental climate, the influx of finely dispersed carbonates into terminal basins enriches bottom sediments with them. During the course of diagenesis, clay and loam-carbonate nodules are formed (Fig. 2b). The sedimentation facies affects the mineral composition of nodules, where the carbonate component consists mainly of calcite. But in

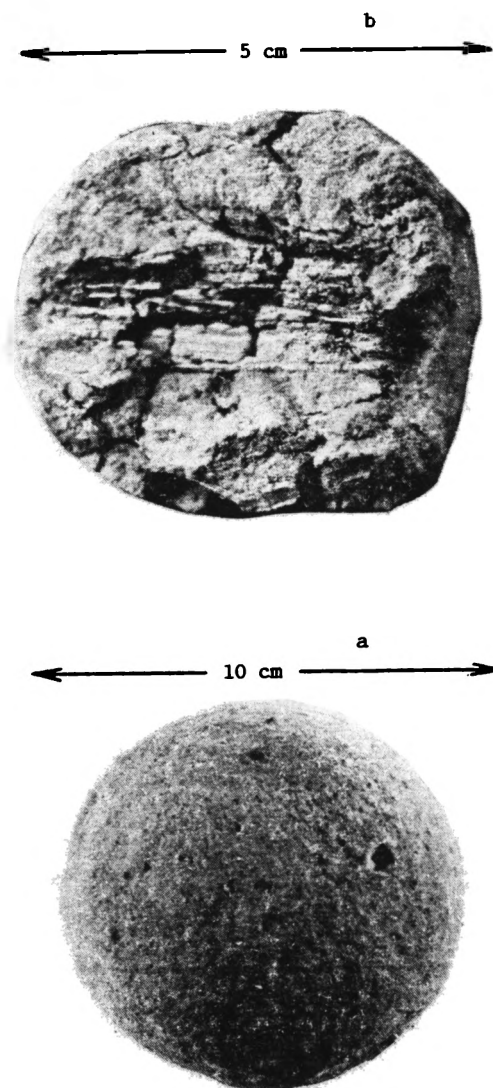


Fig. 2. Diagenetic nodules from glacial-marine sediments: a) internal structure of a nodule consisting of clastic matter and iron sulfides forming a pseudomorphosis after a vegetable remain at the center; b) spheroid nodule consisting of clastic matter cemented with carbonate.

marine rocks there is a high admixture of iron (Fe_2O_3 up to 5.76%; FeO up to 11.98%), manganese (MnO up to 7.8%), and magnesium (MgO up to 7.58%). By contrast, nodules from brackish and freshwater clays and silts have these maximum concentrations (%): Fe_2O_3 4.62; FeO 1.95; MnO 2.64; MgO 4.86 [1, 2].

Freezing and underground ice formation (cryogenesis) determine the specific evolution of sedimentary rocks in the cryolithozone and results in various types of cryogenic structure of permafrost sediments. They vary substantially depending on the general lithogenesis stage and facies at which the freezing took place: during sedimentogenesis (in subaerial and sub-aquatic settings) and during diagenesis (in bottom conditions in the littoral zone of final discharge basins), or after completion of the set of diagenetic processes and emergence from under the sedimentation water level of well-lithified rocks, i.e., during epigenesis in the broad sense (Fig. 3). Depending on the time when sediments pass into a frozen state three freezing mechanisms are distinguished against the general background of lithogenesis: syngenetic, diagenetic, and epigenetic.

Cryolithogenesis

(Sedimentary rock formation in the cryolithozone)

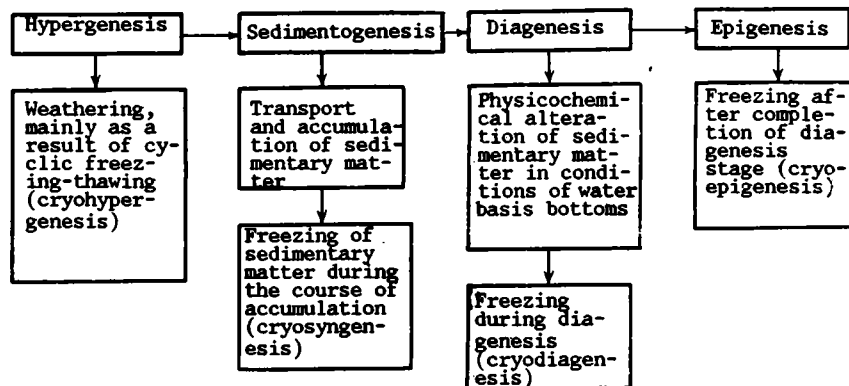


Fig. 3. Stages of cryolithogenesis as the process of sedimentary rock formation.

Syngenetic freezing is typical for slow-flood valley, deltaic, laida, and certain other types of sediments. Frozen beds are built up as sediment accumulates. High-ice sediments with frequent ice interlayers of various thickness (from 0.5 to 7 cm) are formed. The process is controlled by layered ice isolation at the contact of the seasonal summer thaw layer and the roof of the permafrost rocks. The general ice content of finely dispersed (sometimes peat-affected) rocks is 70-80 or even 90% of the rock volume. This is a peculiar cryogenic variety of alluvial-flood valley, deltaic, laida, and slope sediments. They cannot be identified with their genetic correlatives outside the cryolithozone, although that is done sometimes. Besides cryogenic structures formed by thin ice layers, these sediments exhibit large vertical (syngenetic) ice veins that permeate their entire thickness and form a polygonal lattice in space. The veins are up to 40-60 m long and 8-10 m wide. Some of the veins are wider than the earth blocks they limit. A combination of the network ice veins and high-ice cryogenic textures drastically modifies the lithologic habitus of sediments (Fig. 4a). Thawing reduces their thickness by a factor of 7-9, and the internal structure retains traces of the former cryogenic state: ground pseudomorphoses after thawed-out ice veins, cryogenic deformations, etc.

Diagenetic freezing in bottom conditions of water basins is caused by cold flows from below and from the side (from the shore). High-ice cryogenic textures are formed, especially sloping lattice-like and network-like textures [2, 5]. The ice saturation of rocks is sometimes so high that, locally, they become ice deposits with thin layers of clastic matter (see Fig. 4b). In this case cryogenesis interrupts and then completes diagenetic alteration of bottom sediments.

The epigenetic mechanism of freezing is observed in marine-shelf and deep-water lake sediments after they emerge from under the sedimentation basin level and after completion of diagenesis. The lithologic appearance is affected least compared with sediments that freeze during sedimento- and diagenesis. Freezing evolves from top to bottom, and film migration of moisture from below occurs in finely dispersed rock varieties. High-ice stratified-reticular cryogenic textures are usually observed only in the sections closest to the surface - to a depth of 5-10 m. Further down the section, approximately 30-40 m, the network of ice streaks becomes less dense. The streaks thicken and then disappear, given way to massive cryogenic textures with ice becoming the cement binding clastic particles. If the freezing strata of basin sediments include aquifers, the picture is entirely different. The high-ice horizons then occur at depths of 100-200 m and more beneath the day surface. Stratal and lens-shaped ground ice is typical for cryogenic strata of basin sediments (marine, lagoon, and lake types). Vertically they measure tens of meters and stretch for hundreds of meters in a lateral direction. Usually, sand interlayers of lenses occur below the ice beds and lenses; the influx of water during the course of rock freezing from the top is provided through these sands.

* * *

In conclusion, it should be emphasized that cryolithogenesis is a special form of sedimentary rock formation that encompasses all of its stages: hypergenesis, sedimentogenesis,

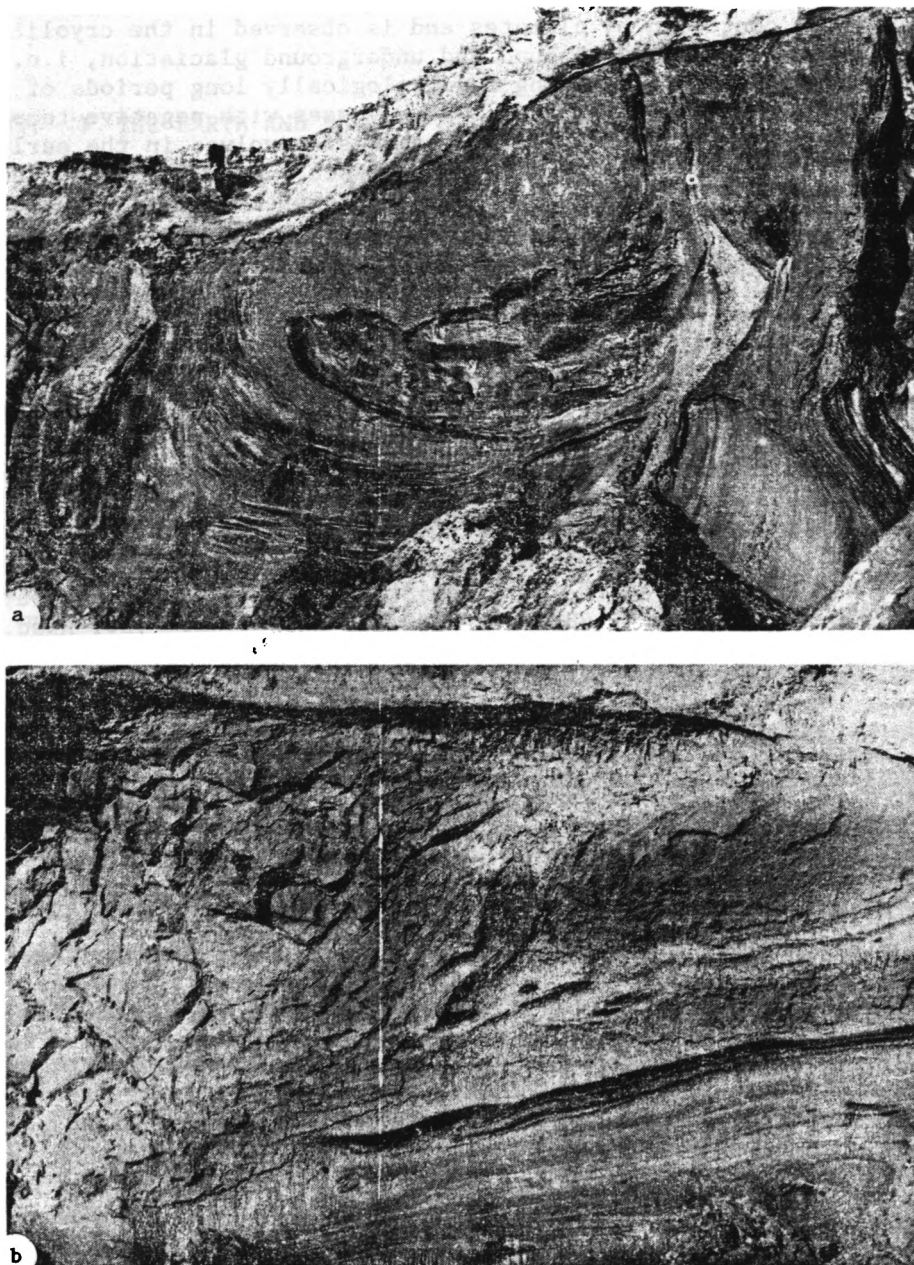


Fig. 4. Cryolithologic structure of two types of sediments that froze according to different mechanisms: a) syngenetically frozen alluvial-deltaic high-ice sand loams permeated with vertical ice veins (cliff height 10-12 m); b) lagoon-marine clays diagenetically frozen in submarine conditions with a sloping lattice of ice interlayers and a lenselike stratified ice placer (frame size 2 x 3 m).

and diagenesis. These processes have their distinctive specifics in the cryolithozone. Cryolithogenesis (freezing and underground ice formation) takes place at different stages of the sedimentary process and largely determines the structural features of sediments formed in the cryolithozone. The facies-genetic type determines the time of appearance of cryogenesis against the general lithogenetic background. These correlations are the main factors to be considered in the permafrost-facies (cryolithogenic) analysis — the main method for understanding the patterns and specifics of cryolithogenesis as a peculiar rock formation process.

Cryolithogenesis is controlled by climates and is observed in the cryolithozone, which includes on continents areas of above-ground and underground glaciation, i.e., glaciers, ice sheets, and permafrost rock strata existing for geologically long periods of time. In seas these are areas of the cryohydrosphere, i.e., water masses with negative temperatures and steady floating ice on the surface [3]. Cryolithogenesis evolved in the early stages of the earth's development in the Late Archean–Early Proterozoic. During geologic history, periods of expansion alternated with times of reduction of these processes in a succession of cryogenic and thermogenic periods. The latest Late Cenozoic cryogenic period began approximately 25 million and probably even 40 million years ago and continues at the present time. It is characterized by widespread occurrence, especially in polar and subpolar regions, of cryolithogenetic products – peculiar sedimentary rocks of various genesis (cryoeluvial, glacial, cryoalluvial, cryolimnic, cryomarine, and others). Studies of the regularities and specifics of cryolithogenesis will shed light not only on the sedimentary process during the latest periods of the earth's history but also on processes in the earth's early stages during the Phanerozoic and Precambrian, when permafrost rock strata and floating ice of coldwater seas, as well as above-ground glaciers, were as prevalent as they are at the present time.

LITERATURE CITED

1. I. D. Danilov, Polar Lithogenesis [in Russian], Nedra, Moscow (1978).
2. I. D. Danilov, Methods of Cryolithologic Exploration [in Russian], Nedra, Moscow (1983).
3. I. D. Danilov, "Cryolithozone of the earth and its subzoning," *Izv. Akad. Nauk SSSR, Ser. Geol.*, No. 1, 12-18, 1983.
4. É. D. Ershov, Cryolithogenesis [in Russian], Nedra, Moscow (1982).
5. E. M. Katasonov, "Types of permafrost beds and cryolithologic problems," in: *Geocryologic and Hydrogeologic Studies in Siberia* [in Russian], Yakutsk. In. Izd-vo, Yakutsk, pp. 5-16 (1972).
6. N. N. Lapina, G. A. Znachko-Yuavorskii, N. N. Kulikov, et al., "The polar-type lithogenesis," in: *Genesis and Classification of Sedimentary Rocks* [in Russian], Nauka, Moscow, pp. 212-217 (1968).
7. O. K. Leont'ev, "On supply of sedimentary matter to the Arctic Ocean," *Vestn. MGU, Ser. 5, Geografiya*, No. 1, 34-38, 1984.
8. A. I. Popov, *Cryogenic Effects in the Earth's Crust (Cryolithology)* [in Russian], Izd-vo MGU, Moscow (1967).
9. N. M. Strakhov, *Principles of the Theory of Lithogenesis*, Vol. 1 [in Russian], Izd-vo Akad. Nauk SSSR, Moscow (1960).
10. N. A. Shilo, *Principles of the Science of Placers* [in Russian], Nauka, Moscow (1981).
11. A. A. Sharbotyan and P. A. Shumskii, "Evolution of the earth's cryosphere," in: *Problems of Paleohydrology* [in Russian], Nauka, Moscow, pp. 143-160 (1976).

The conventional criteria determining aquatic sedimentary origin are shown to be untenable for Early Archean gneiss complexes as they are inconsistent with the general high-temperature environment of the Early Archean. Early Archean gneiss areas may have been formed as a result of heat-caused disintegration of the ancient igneous rocks on the earth, with some participation of chemogenic processes.

Sedimentation mechanisms in an aquatic medium in the most ancient geologic epochs have been debated since the nineteenth century. Until recently, the problem was solved in the spirit of uniformity – the assumption that the only applicable method is a perfect analogy between ancient and recent sedimentation processes. Although uniformism as a geological theory was at variance with some physical findings as early as in the late nineteenth century and is philosophically unacceptable to an investigator relying on dialectical materialism [6], analogies, especially perfect analogies which are the most conservative component of knowledge, have become popular, because of their elementary simplicity, in studies of Early Archean formations and processes, such as the notion of an eternal and invariable world ocean, a transfer of plate tectonics to the Early Archean, analogies in petrochemical and geochemical descriptions of Early Archean rocks with petrochemistry of recent sedimentary rocks or the petrochemistry and geochemistry of young intra- and periocenic structures, etc.

The first critique of the methodology of uniformism was attempted in the 1960's and early 1970's, when scientific progress revived the notion of a primary crust. The nature of Early Archean gneisses that were interpreted by uniformists as sediments of the primary ocean also came under scrutiny. This evoked vigorous resistance on the part of orthodox uniformists led by Sidorenko, who asserted categorically that sedimentation from the Archean up to the Recent type evolved in similar conditions and that fundamental sediment-forming processes remained largely invariable [17, 18]. The philosophy of Early Precambrian and especially Early Archean studies until recently relied on the archaic principle of uniformism combined with Neptunist theories. Today, no one can be satisfied with this primitive state of the methodology of studies of the Early Precambrian. Major discoveries in the geology of the Precambrian over the last five decades and breakthroughs in related fields compel one to revise obsolete methodology and to develop modern methods based on recent findings of the general regularities of the structure of Precambrian regions and the basic development trends of geological processes during the Early Precambrian [8].

The type of crust formation, the physicochemical conditions on the planet's surface, and the patterns of rock-forming processes during the Early Precambrian can now be divided into two epochs: 1) formation and primary consolidation of the crust (Early Archean); 2) evolution of geological processes on a stabilized crust (Upper Archean and Lower Proterozoic). The geological borderline between these two types of rocks is traced along the first greenstone belts. A nominal chronological boundary lies at approximately 3.5 billion years. The first epoch covers all pre-greenstone rocks of shields (ancient granitoids, gneiss, and granulitic strata); during the second epoch, rocks formed on the surface of the crust (supercrustal rocks) appear and become predominant. The border between these two epochs marks a drastic qualitative change after primary consolidation of crust toward exogenous processes, including initial aquatic sedimentation (see Fig. 1). The principal geological characteristics of these two epochs are so dissimilar that the methodologies of studying them should also be different.

A modern methodology for study of the Early Archean should be based on the obvious fact that the early geological stage in the development of the earth's crust is closely related to the pregeological period of the planet's development, is its direct continuation, a special stage that has no analogs in the subsequent history of the earth. Khoreva [22] described it

Institute of Geology and Geochronology of the Precambrian, Academy of Sciences of the USSR, Leningrad. Translated from *Litologiya i Poleznye Iskopaemye*, No. 1, pp. 13-19, 1990. Original article submitted February 16, 1989.

as a late planetary stage in the formation of protocontinental crust. A reliable description of the pregeological epoch of the earth's shells is crucial for proper understanding of the geological processes that formed the ancient earth's crust.

Regrettably, identifying any direct cosmic relicts and cosmogeological structures in the ancient earth's crust, that has been transformed many times after its origination, is impossible. There are no other planets in the solar system similar to the earth; especially, no other planets have the specific feature of the earth's crust of a granite layer. As a result, direct analogies comparing the earth to the moon or Venus add virtually nothing to an understanding of the origins of this layer. The evolution and primary crust on the earth and on other planets diverged early on in planet formation, and comparison and analogies are extremely limited. A closer look may reveal some approximate and obscure geological analogies, but nothing sufficient for solving the problem as a whole.

Modern theories of the origin of planets of the solar system based on astronomy, cosmic chemistry, the physics of space processes, and studies of space matter and deep earth rocks provide much more information for understanding the geology and genesis of the Early Archean. All modern cosmogenic theories assume that planets were formed by accretion from a protoplanetary gas/dust cloud (nebula) and vary in type of accretion (homogeneous, heterogeneous, hot, cold, fast, and slow) and the composition of the accreted matter. The compositions of accretion products and gas/dust condensates varied in the different portions of the nascent crust, which was responsible for the lateral inhomogeneity of the crust as well as the basic properties. The ancient geological processes had to deal with the matter provided by accretion rather than with sediments of the "primary world ocean."

The reworking of dumped primary accretion products by intracrustal processes did not substantially modify the composition of laterally inhomogeneous segments. All the Early Archean blocks identified today represent originally accretional masses of various composition reworked in mixes of various products - igneous (plutonic and volcanogenic), disintegrational, and chemogenic. Of course, no unaltered accretional products have survived. Their role in the formation of the primary crustal shell is retained in "geologic memory." In particular, cherts and enderbites of the Antarctic have been found to contain ^{207}Pb concentrations anomalous or earth rocks or meteorite ratios of lead isotopes [19]. The isotopic composition of sulfur from rocks of the Iengrian series of the Aldan Shield matched meteoritic composition [3, 4]. Vinogradov notes that zero composition (i.e., meteoritic) should be characteristic of sulfur at any point on the earth during the prebiological stage of its history.

Studies on Precambrian shields have determined that bipyroxene crystalline schists are the most ancient on earth. They have no analogs among sedimentary rocks and are igneous formations. These are the earliest products of reworking of accretional material. A high temperature is needed to crystallize igneous rocks, which suggests a certain period of heating. The time of maximum heating is set at approximately 4 billion years ago, it separates the cosmic history of the earth's development from its geological history.

In conditions of planetary cooling that occurred at high temperatures which ruled out formation of water in a liquid phase, the future geostructural regions (areas) began to be formed. They consisted of magmatogenic entities whose chemical composition was inherited from the original accretional masses and the mineral compositions were determined by the temperature regime of the processes. Two types of such primary crustal areas are distinguished: granulite-enderbitic and tonalitic.

The White Sea, Kola, Bug, Upper Aldan, and other series are geologically related to primary crustal areas; these are rocks which on a scorching hot and waterless planet, could only be formed by high-temperature disintegration of primary crustal enderbites and tonalities with some contribution of chemogenic processes. Disintegration products are interstratified with volcanites, which also indicates high temperatures. These polygenous early crustal gneiss complexes constitute the second group of Early Archean formations.

According to traditional methodological principles based on uniformism and the hypothesis of a primary water ocean, the early crustal gneiss complexes were classified as sedimentary rocks (from a water medium). The new principles of the initial history of the earth, directly related to cosmic history and ruling out any primary water ocean, make it necessary to revise the old theory of Early Archean sedimentation. To begin with, we must reconsider the validity of the methods of study and the criteria for the water-sediment origin of strongly metamorphosed rocks which have lost the primary state of aggregation because of metamorphism and have not preserved their original textural and structural characteristics.

glomerates, with far-reaching implications. Speculations are often based on large-scale striation, which is equated with rhythmicity. The members of the rhythms are taken to be packets of tens or hundreds of meters, and the section is artificially constructed as a set of so-called rhythms.

Striation of any scale is a polygenetic phenomenon. It is a natural geological state of metamorphic rocks of any genesis. With the current state of our knowledge, only rhythmic gradational stratification can be an indication of the sedimentary origin of a rock; no such structures are reliably known in Early Archean complexes.

Rounded zircons have been described as indicators of sedimentary rocks. It was assumed that they were rounded in a water environment. By exact goniometric measurements, Yudina demonstrated that such "rounded" zircons in granulites are polyhedra with underdeveloped facets [22, 23]. Sonyushkin [20] determined by studies in a raster electron microscope the absence of signs of an exogenous effect on zircon crystals (scratches, burrows, or traces of alkaline dissolution) in amphibolitic and granulitic rocks. Zircon in rocks of these facies is metamorphogenic; the rounded shape of some of the grains cannot be viewed as an indication of the sedimentary origin of the rock. After Sonyushkin's studies the problem of clastic zircon in highly metamorphosed Early Archean complexes is no longer a legitimate subject for discussion.

The presence of carbon (graphite, carbonate, and other forms) believed to be of a biogenic origin is considered an indicator of sedimentary rocks. Signs of primitive life, even if they can be established in early crustal complexes, do not as such prove a marine environment. Lozina-Lozinskii [13] proved the possibility of life on each evolving on dry land outside an aquatic environment. Modern bacteria have an extreme vitality. They can survive any temperatures and a strong reactive environment and can live even in nuclear plants. Owing to the capacity of organisms to pass to anabiosis, no unfavorable conditions (such as ultraviolet radiation, dryness, or abrupt temperature changes) could prevent development of life on land. Anabiosis and hypobiosis (massively suppressed life activity), typical for lower forms of life and at the cell level, can be evidence that was formed on land [13]. For single-celled and primitive life forms carbon could be inherited in abundance from carbonaceous chondrites of the accretional stage. Besides, Chernyshev et al. [24] have demonstrated that no methods of C_{org} determination in rocks are reliable. There are many mechanisms of graphite formation by an abiogenic route through a series of metamorphogenic reactions during endogenous processes; this possibility has been studied, for example, by Bukharev [2] on graphite-containing gneiss of the Podolye block of the Ukrainian Shield.

Carbonate rocks (marbles, calciphyries, etc.) were traditionally viewed as sedimentary rocks (e.g., [11]); their presence was seen as an indication of a marine environment. A detailed study of the geological characteristics of carbonate rocks in the Aldan Shield shows that they often form cutting and symmetrically zonal bodies containing fragments of gneiss and skarn inclusions, which are evidence of a chemical interaction of silicate and carbonate material, i.e., they are not sedimentary beds [10].

Many cardinal problems remain unresolved. In particular, common rocks such as biotite, hornblende, and other gneisses, that occupy most of the Early Archean sections in many locations, have yet to be interpreted in genetic terms.

Generally, none of the above criteria provide an unambiguous answer. Geological reconstructions built on such foundations are precarious. Classifying a rock as a terrigenous sediment formed by deposition from an aquatic medium does not rely on any proofs and is inconsistent with the general environment in primary crustal foundation regions: high-temperature chemically active medium and subaerial disintegration where a water ocean could not exist. The geology and rock formation conditions during the Early Archean remain unclear in many respects, but even now it is obvious that neither uniformism nor actualism can explain the processes of initial formation and primary consolidation of the crust. A new approach is needed for understanding the genesis of early crustal gneiss complexes.

The first steps in this direction have been taken. In particular, Bukharev [2] has made an innovative study of the genesis of the rock of the Podolye granulite-enderbite area of the Ukrainian Shield. He interprets hypersthene gneiss as rocks formed by metamorphism of volcanoclastics and its redeposition products during active volcanic events rather than as rocks formed from basic effusives (lava streams, covers, etc.). Quartzites are attributed to a volcanogenic-chemogenic genesis due to mass transport of silica by volcanogenic hydro-

thermal solutions (similar to Nakovnik's "secondary quartzites"); formation of high-alumina gneiss is attributed to alumina accumulation during fumarole-solfatara decay of basic volcanites. It appears likely, however, that high-temperature disintegration of volcanites was the predominant process in a chemically active aerial medium of the planet's external shell. These processes were not directly related with the activity of the volcanic apparatus. Products of subaerial disintegration are confirmed by modern physical and chemical weathering on Mars, Venus, and other planets.

The real study of the genesis of early crustal rocks is yet to be done. Structural analysis associated with subtle geochemical, petrographic, and microtextural investigations acquires primary importance for understanding the true origin of rocks. In the coming decade it will be important to develop a collection of descriptions and drawings of structures studied in minute detail as the basis for investigating and determining the genesis of Early Archean rocks.

At the boundary between the Early and Upper Archean, when the consolidated basement was formed, an epoch of geological processes on stabilized crust was begun, which continues now. New geological factors and phenomena appeared at that boundary: the first clear signs of sediment accumulation from the aquatic medium (e.g., [16]), intrusions that penetrated into the rigid frame, central shield volcanos, and zonal metamorphism caused by intrusions or local heat flows. A basic transition to a germanotypic environment takes place.

All subsequent geological history from the Late Archean to recent times followed the law of irreversible evolution. This law is based on a historic interpretation that is diametrically opposed to the principle of uniformism. A theory consistently constructed along a historic line was defined by Leonov [12] as scientific evolutionism. The present author believes that, with regard to the geological environment the concept can be redefined as geological historicism or a concept of evolutionary development of geological processes.

During the epoch of irreversible evolution - from the Late Archean to the Neogene - the outer shell of the earth and exogenous processes evolved in a direction of increasing intensity of certain processes and quantitative expansion of their material carriers. This line of evolution is exemplified by the birth at the border between the Early and Late Archean of a surface hydrosphere and the gradual expansion of its quantity during geological history (see Fig. 1). At the same time, the sedimentation areas increased from lake-like small basins of the greenstone belt time to shallow intracratonic basins during the Early Proterozoic time to epi- and intercontinental seas and recent oceans [21]. A description of sedimentation during the epoch of irreversible evolution should be the subject of an independent extensive analysis.

LITERATURE CITED

1. I. A. Bergman, "Applicability of petrochemical methods to reconstructions of the sedimentary Precambrian," *Litol. Polezn. Iskop.*, No. 6, 110-122 (1987).
2. V. P. Bukharev, "Reconstruction of the primary composition of supercrustal rocks in the Podolye megablock of the Ukrainian Shield," *Geol. Zhurn.*, 47, No. 5, 101-110 (1987).
3. V. I. Vinogradov, "Historicism in geochemistry in light of new data on the isotopic composition of sulfur," in: *Essays in the Geochemistry of Elements* [in Russian], Nauka, Moscow, pp. 244-274 (1973).
4. V. I. Vinogradov, "Isotopic composition of elements, mantle degassing, and development of the earth's gas-water shell," in: *Earth Degassing and Geotectonics* [in Russian], Nauka, Moscow, pp. 23-30 (1980).
5. I. I. Vishnevskaya, "Disthene quartzites of Kola series of the Precambrian (Kola Peninsula)," *Izv. Vyssh. Uchebn. Zaved. Geol. Razved.*, No. 6, 45-51 (1988).
6. B. P. Vysotskii, "Development of actualism as a scientific method in geology," *Ocherki Istor. Geol. Zn.*, Issue 8, 64-103 (1959).
7. A. S. Efron, A. S. Voinov, and V. V. Gordienko, "Geochemical characteristics of metamorphic rocks of the White Sea complex," in: *Micaceous Pegmatites in Northern Karelia* [in Russian], Nedra, Leningrad, pp. 43-95 (1976).
8. A. N. Kazakov, "Methodology of geologic studies of the early stages of geologic history," in: *Tectonics and Mineral Resources of the Precambrian in Siberia and the Far East (Conference Proceedings)* [in Russian], Irkutsk, pp. 85-86 (1988).
9. M. D. Krylova, A. N. Kazakov, and F. P. Mitrofanov, *Conglomerates and Pseudoconglomerates of the Precambrian* [in Russian], Nauka, Leningrad (1988).

10. A. L. Kulakovskii and N. N. Pertsev, Allochthonic carbonates of the Precambrian in Central Aldan," *Izv. Akad. Nauk SSSR, Ser. Geol.*, No. 1, 52-68 (1987).
11. E. A. Kulish, *Sedimentary Geology of the Archean in the Aldan Shield* [in Russian], Nauka, Moscow (1983).
12. G. P. Leonov, "Historicism and actualism in geology," *Vestn. MGU, Ser. Geol.*, No. 3, 3-15 (1970).
13. L. K. Lozina-Lozinskii, "Where did life first arise: In the ocean or on Land?" *Zh. Evol. Biokhim. Fiz.*, 21, No. 3, 303-307 (1985).
14. A. N. Neelov, *Petrochemical Classification of Metamorphosed Sedimentary and Volcanogenic Rocks* [in Russian], Nauka, Leningrad (1980).
15. A. A. Predovskii, *Geochemical Reconstruction of the Primary Composition of Metamorphosed Volcanic-Sedimentary Precambrian Formations* [in Russian], Kirovskii Rabochii, Apatity (1970).
16. V. A. Rudnik, "Zonality and composition of the earth's geosphere," in: *Development of A. N. Zavaritskii's Ideas in Modern Petrology* [in Russian], Nauka, Moscow, pp. 57-89 (1986).
17. A. V. Sidorenko, "On a unified historic-geologic principles in studies of the Precambrian and its bearing on the pre-Paleozoic history of the earth," *Dokl. Akad. Nauk SSSR*, 186, No. 1, 166-169 (1969).
18. A. V. Sidorenko, "Precambrian sedimentary geology and implications for the earth's Pre-paleozoic history," *Sov. Geol.*, No. 2, 3-16 (1975).
19. E. V. Sobotovich, E. N. Kamenev, A. A. Komaristyi, and V. A. Rudnik, "Ancient rocks of the Antarctic (Enderby Land)," *Izv. Akad. Nauk SSSR, Ser. Geol.*, No. 2, 3-16 (1975).
20. V. E. Sonyushkin, "Zircon of metamorphic rocks in a scanning electron microscope: Genetic crystal morphology and anatomy," *Preservation of the Composition, Structure, and Texture of Sedimentary and Volcanogenic Rocks in Metamorphism* (Proceedings of a Conference on Geological and Geochemical Methods of Reconstruction of Primary Nature of Metamorphic Rocks) [in Russian], Preprint of the Leningrad State University, Leningrad, pp. 17-18 (1988).
21. P. P. Timofeev, V. N. Kholodov, and V. P. Zverev, "Evolution of the mass of natural waters of the earth and the sedimentary process," *Dokl. Akad. Nauk SSSR*, 288, No. 2, 444-447 (1986).
22. B. Ya. Khoreva, "Structural and constitutional specifics and formation mechanisms of Archean granulite-gneiss complexes," in: *Minerals, Rocks, and Mineral Deposits in Geologic History* [in Russian], Nauka, Leningrad, pp. 161-171 (1985).
23. B. Ya. Khoreva, *A Map of Metamorphic and Related Granitoid Formations in the Territory of the USSR, Scale 1:10,000,000* [in Russian], VSEGEI, Leningrad (1986).
24. V. G. Chernyshov, O. A. Chernyshova, and O. V. Vershkovskaya, "On methodological errors in analysis of organic matter in rocks," *Dokl. Akad. Nauk SSSR*, 292, No. 6, 1472-1476 (1987).

ISOTOPIC COMPOSITION OF SULFIDIC SULFUR OF THE ZAZHOGIN SHUNGITE
DEPOSIT (SOUTH KARELIA)

G. V. Shatskii

UDC 550.4:552.57(470.22)

Pyritic mineralization localized in shungites, shungitized metasediments, and, to a lesser extent, in gabbro diabases is studied. The isotopic composition of pyrite sulfur clearly indicates a crustal source of sulfidic sulfur.

The diverse properties of shungites of various chemical and mineral compositions make them suitable for various uses as a substitutes for coke, graphite, construction material in production of silicic ferroalloys, cast-iron, and nonferrous metals [17]. Shungite beds are a potential source of sulfidic mineralization.

The sources of sulfur have remained unclear until now. There were two main theories. Some authors [5, 7] attributed accumulation of large quantities of carbonaceous matter to basic magmatism, which produced covers and sills of diabases that take up a considerable portion of the section of shungite-bearing rocks. Pyrite metallization is interpreted as resulting from exhalational activity of basaltoid magmatism. An alternative view sees sulfides (most of them) and the enclosing shungites as sedimentary formations unrelated to magmatism.

The main objective of our isotopic study was to determine the source of the sulfur from sulfidic mineralization. There are significant differences between the isotopic characteristics of deep (juvenile) sulfur and those of sulfur that has experienced the sedimentary cycle. It is also important to estimate potential variations in isotopic ratios caused by metamorphism, because rocks in this region have passed through metamorphism of greenschist facies.

Shungite rocks are localized in the Onega trough. It is an irregularly shaped oval structure stretching from Petrozavodsk to Medvezhegorsk for 130 km; with a width of 120 km, it has an area of over 10,000 km² [6].

Shungite-bearing rocks belong to the Trans-Onega suite, consisting of two subsuites: the lower, originally sand-clay-volcanogenic subsuite, and the upper subsuite, the true shungite-containing formation, up to 600-650 m thick [14].

In turn, three members are distinguished in the upper subsuite. In the Zazhogin deposit, shungite beds are confined to the second member of the subsuite (Fig. 1), 200-250 m thick. The lower part of the member is made up of shungite-bearing metatuffosiltstones, metasiltstone, shungitic quartz-sericite-biotitic rocks (metapelite and metasiltstone), and dolomeites [10]. Up to the section, psephitic tuffs are interstratified with shungitic calcareous tuffites and layers of shungite-bearing dolomites.

The next portion of the section consists of tuffosiltstone and tuffosandstone, shungite-containing calcareous tuffosiltstone and tuffosandstone, shungite-containing calcareous tuffosiltstone, and quartz-sericitic fine crystalline schists; up the section, there are conformable diabase bodies and, importantly, beds of shungitic rocks with carbon contents from 19 to 55%. Lyddite-shungite-dolomitic formations are at the top of the member.

There are several classifications of shungitic rocks. Borisov's classification [1] is most common. There are five types, depending on carbon content: shungite I (above 98%), shungite II (35-75%), shungite III (20-35%), shungite IV (10-20%), and shungite V (less than 10%). Structurally, three types of shungitic rock are distinguished: massive, striated, and brecciated [15]. Rocks of the Trans-Onega suite have experienced regional metamorphism of the greenschist facies [8]. Shungitic strata, like most black schist formations, have a heightened sulfidic content. The amount of sulfides in various types of shungites ranges

Geological Institute, Academy of Sciences of the USSR, Moscow. Translated from *Litologiya i Poleznye Iskopaemye*, No. 1, pp. 20-28, 1990. Original article submitted June, 10, 1988.

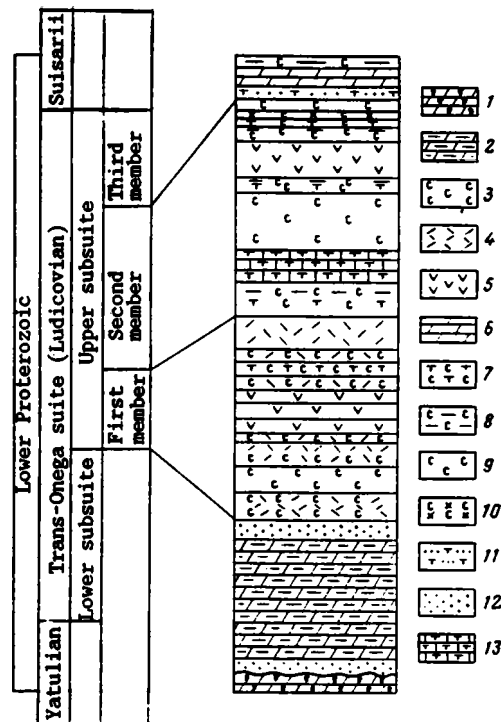


Fig. 1. Lithologic-stratigraphic column of Maksovo-Zazhagin section (after [15]): 1) carbonaceous rocks of the Upper Yatuli-an; 2) micaceous dolomites of the lower subsuite; 3) shungitic rocks with albite-chloritic mineral base; 4) limestones with actinolite; 5) igneous rocks of a basic composition; 6) shungite-bearing dolomite; 7) shungitic tuffs and tuffites; 8) shungitic rock quartz-sericite-biotitic schists; 9) shungitic rocks with quartz-sericite-biotitic base; 10) lyddites; 11) shungite-bearing tuffosiltstones and sandstones; 12) sedimentary rocks of the terrigenous series; 13) tuffogenic limestones.

from fractions of a percentage point to 20%. Furthermore, there are isolated sulfide lenses 10-15 cm thick. In sections with diverse structural shungite types sulfidic mineralization is distributed unevenly. Most sulfides are concentrated in striated varieties. There are fewer sulfides in massive rocks. Brecciated rocks have minimal amounts of pyrite. The set of sulfides is extremely uniform: pyrite, chalcopyrite, sphalerite, pyrrhotite, marcasite, melnikovite, and kubanite (?). Over 95% of the ore mineral consists of pyrite.

Structurally, three groups of sulfidic mineralization are identified: lens-massive, inset, and vein types. The former two groups are widespread. Among disseminated structures there are uniformly disseminated, banded, and schlieren-nodular. Schlieren segregates are found as single units and combined with fine pyrite insets. There are two types of veins: isolated veins (1-3 cm) and networks of thin branching veinlets. Pyrites of disseminated textures are represented by two generations. They vary in shape, size, and spatial relationship to non-ore minerals. Pyrite-I: size from 0.005 to 0.05 mm, allotriomorphic, less often hypidiomorphic texture. Pyrite of the first generation is often found in fine coalescence with quartz and carbonaceous matter. Most mineral isolates are isometric. Pyrites of the first generation include a considerable number of globular and small sperulite framboids. Pyrite-I consists mostly of uniform and banded structures.

Pyrite-II consists of idiomorphic, mostly cubic crystals, from 0.05 to 2 cm. The crystalline texture is often porphyroblastic, representing metamorphogenic crystallization. Large crystals have skeletal forms and relicts of the parent rock entrapped during metamorphism and enclosed in metacrystals. Sedimentary structures are revealed by etching pyrite of second generation. Correlations of the contents of sulfidic sulfur and free carbon indicate no functional relationship. One can only say that the higher-carbon shungite varieties contain less sulfides than do calcareous striated varieties.

Two types of pyrite mineralization are distinguished genetically. The first type includes the presumably primary sedimentary syngenetic sulfide; the second type consists of metamorphogenic-hydrothermal pyrite entities [6].

The sedimentary type is represented by pyrites with inset, nodular, and lens-massive structures. This pyrite has been recrystallized partially or completely "in situ" during regional metamorphism. In some locations relict structural-textural features of sedimentary mineralization are observed: framboidal pyrite, fine globular pyrite, and certain collomorphic structures of larger nodules and concretions. Insets are often stratified, conforming to the layers of metasedimentary rocks.

The metamorphogenic-hydrothermal type consists of pyrite veinlets of various kinds.

Isotopic analysis was performed at the Laboratory of Isotope Geochemistry and Geochronology of the Geological Institute of the USSR Academy of Sciences.

Rocks with pyrite were ground and sifted to a fraction of 0.5-1 mm. The pyrite monofraction was selected from this fraction. The sulfides were converted to SO_2 by the conventional method. The gas was analyzed on an MI-1201V mass spectrometer. The values of $\delta^{34}\text{S}$ are expressed in promille points with respect to the Sikhote-Alin meteorite standard. The reproducibility of $\delta^{34}\text{S}$ is $\pm 0.5\%$.

Results of Isotopic Analysis. There is a wide scatter of $\delta^{34}\text{S}$, amounting to 53.4%. Most values are in the positive region (of 51 analyses, 7 were negative). Over half the values are in the range of (+15-+22) (Tables 1 and 2, Figs. 2 and 3).

The widest range of $\delta^{34}\text{S}$ values is observed in rocks with higher carbonate content: calcareous metatuffites, metasiltsstones, shungites, and dolomites. All negative values of metasedimentary rocks belong to this group. The true shungites have a narrower range. All the values of $\delta^{34}\text{S}$ are positive with a scatter 18.6%. Basic igneous rocks have a minimal data divergence: at most 1.5%. Considering reproducibility, one can conclude that the isotopic composition is nearly homogeneous. The isotopic values of pyrites of all the structural-textural groups show no considerable diversity. One can only say that disseminated and schlieren-like nodular forms are slightly less homogeneous, although the number of analysis in each group was not the same.

We attempted to correlate the isotopic composition with the content of C_{org} in SiO_2 in the rock. The correlation for C_{org} was unclear. With increasing C_{org} , the isotopic composition of sulfides became heavier (Fig. 4). However, because of the small number of paired determinations of C_{org} and $\delta^{34}\text{S}$, this correlation was not reliably established.

For the entire section of the deposit and within members, the isotopic ratios of pyrite sulfur varied with no significant correlation with hypsometry (Fig. 5). This is also true of the distribution of $\delta^{34}\text{S}$ in the lateral series.

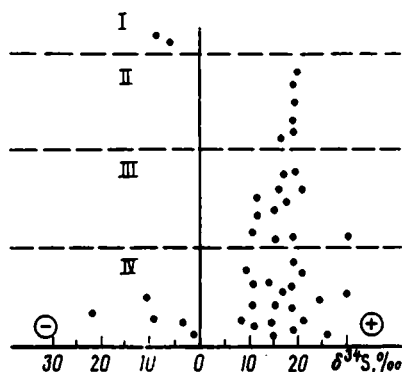


Fig. 2

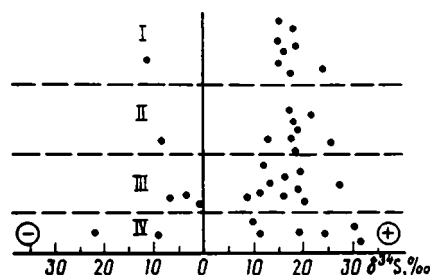


Fig. 3

Fig. 2. Distribution of $\delta^{34}\text{S}$ of pyrites according to lithologic rock types: I) metasomatite, lyddite; II) metadiabases and metabasalts; III) shungite varieties ($\text{C}_{\text{CB}} > 10\%$); IV) calcareous metatuffites, metasiltsstones, shungites ($\text{C}_{\text{CB}} < 10\%$), carbonaceous dolomites.

Fig. 3. Distribution of $\delta^{34}\text{S}$ of pyrite according to morphologic mineralization type: I) veins; II) lens-massive; III) disseminated; IV) schlieren-nodules.

TABLE 1. Distribution of Pyrite $\delta^{34}\text{S}$ According to Lithologic Rock Types

Characteristics of surrounding rocks	Range of values
Calcareous metatuffites, metasilstones, shungites ($C_{CB} < 10\%$), carbonaceous dolomites (28)	$-22,1 \div +31,3$
II-IV shungite varieties ($C_{CB} > 10\%$) [14]	$+11,9 \div +30,5$
Metadatabases and metabasalts (6)	$+17,8 \div +19,2$
Metasomatite (?), lyddite (2)	$-6,9; -9,4$ (respectively)

TABLE 2. Distribution of Pyrite $\delta^{34}\text{S}$ According to Morphologic Mineralization Type

Textural-structural type of pyrite mineralization	Range of values
Veins (12)	$-11,9 - +23,7$
Lens-massive (9)	$-8,3 - +25,4$
Disseminated (14)	$-6,9 - +27,2$
Schlieren-nodules (8)	$-21,1 - +31,3$

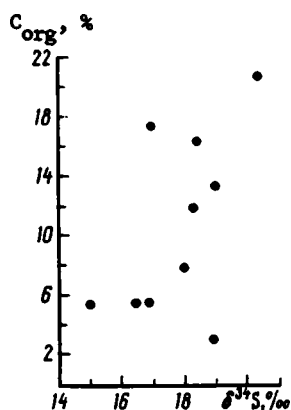


Fig. 4. Values of $\delta^{34}\text{S}$ vs. free carbon content.

The following investigations were performed to evaluate the role of regional metamorphism in the redistribution of sulfur isotopes in sulfides. Cores were taken that contained several types of pyrites in terms of morphological data and genesis/formation type: disseminated-inset, cutting vein-type, and recrystallized coarse crystalline granoblastic pyrites. Pyrite monofractions were isolated from the cores. Such core included different genetic types of pyrite mineralization and were taken from three different parts of the section. The differences in isotopic composition of these specimens did not exceed 3%. (specimen) 1a + 18.9; 1b + 19.2; 2a + 17.5; 2b + 21.3; 3a + 24.3; 3b + 27.2. These data clearly show that the differences in isotopic composition of premetamorphogenic scattered pyrite mineralization and metamorphogenic scattered pyrite mineralization and metamorphogenic hydrothermal-vein type mineralization are insignificant.

The fact that a wide scatter of $\delta^{34}\text{S}$ values and a lack of a "system" in the spatial distribution of isotopically different pyrites is not characteristic of the metamorphogenic isotopic distribution indicates that regional metamorphism had no important impact on the isotopic composition of sulfidic mineralization. Furthermore, there is no substantial redistribution as a result of mobilization and redeposition of ore material if all transport occurs within a few dozen centimeters.

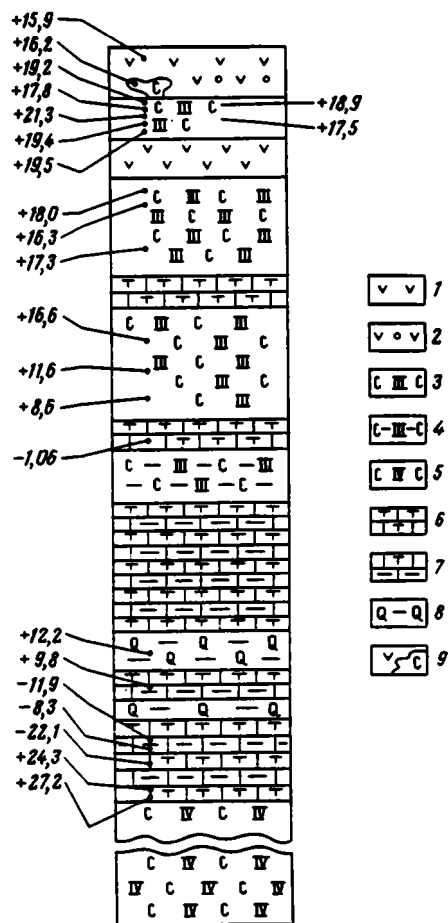


Fig. 5. Distribution of isotopically characterized pyrite in the section of bore hole 224: 1) metadiabase; 2) metadiabase of amygdaloid-stone structure; 3) shungite III; 4) shungite III (banded); 5) shungite IV; 6) calcareous metatuffite; 7) interstratification of calcareous metatuffite and metasiltstone; 8) silicic-carbonaceous schists; 9) shungite xenoliths in metadiabases.

Discussion. The basic reference standard in sulfur isotopic studies is sulfur from meteoritic troilite (Diablo Canyon and Sikhote-Alin in the USSR). Many of the basaltoid and ultrabasic rocks, believed to be the deepest formations and with some reservations considered melts of mantle matter, have isotopic sulfur compositions similar to that of meteorites [11]. The average isotopic composition of the earth's sulfur seems to match meteorite and the deepest (mantle) distributions, i.e., it is close to zero. An isotopic composition similar to that of the mantle ought to have existed also in the oldest subsurface rocks of the earth that were formed before sulfur circulation developed, i.e., before life and free oxygen appeared on earth.

Sulfur in the earth's crust has wide variations of $\delta^{34}\text{S}$ (from -70 to +70%). Seawater sulfate is the main source of isotopically heavy sulfur. Its $\delta^{34}\text{S}$ value is close to +20%. As has been shown in [3], it is a kind of planetary constant that is descriptive of the sulfur balance in the sedimentary cycle and effectively provides a second reference point for isotopic geochemical reconstructions.

The dispersion of isotopic ratios is associated with the fact that during sulfate reduction sulfur isotopes become divided: reduced forms become depleted of the heavy isotope, making the residual sulfate relatively more enriched in it. Bacterial sulfate reduction is a powerful engine of isotope separation in natural environments. Broad variations of $\delta^{34}\text{S}$ in shungite-bearing rocks show that this mechanism was fully realized already in the Early Proterozoic. The averages of +15 to +20% are typical for the isotopic composition of sulfur in re-

cent marine sulfate ions. Sulfur of the sedimentary cycle thus clearly contributes to development of sulfide mineralization of the Zazhogin deposit. Isotopic data also indicate that the igneous rocks of the shungite-schist section in some form derived sulfur from surrounding rocks: pyrite of diabases has a high $\delta^{34}\text{S}$ value in an extremely narrow interval of +17.8 to 19.2%. Such an isotopic composition matches the average $\delta^{34}\text{S}$ of pyrites from the surrounding rocks.

As mentioned earlier, regional metamorphism had no significant effect on the isotopic composition of sulfur, which apparently reflected the environments of sediment lithogenesis. Information derived from studies of younger rocks appears to be a legitimate source of information for understanding these environments.

Sulfides formed during sedimentogenesis and early diagenesis usually have negative values (an average of -15 to -20%). With an unlimited source of sulfate from seawater ($\delta^{34}\text{S} = +20$) and at low temperatures, the isotopic effect from bacterial sulfate reduction typically amounts to some 40%. When sulfate availability is limited (for instance, during later stages of diagenesis when water exchange between bottom water and sediment may be discontinued), residual sulfate solution can become gradually of a heavier type with continued reduction [3]; sulfides with heavy sulfur up to values of +20% or more can be formed as a result. Yet, the volume of such sulfides is extremely small. For instance, to obtain a sulfate residue enriched in the heavy sulfur isotope to 50% (at a fractionation factor of 1.025) [20], as much as 99.999% of the initial sulfate has to be reduced. This means that from 1 t of initial sulfate just 100 mg can be obtained with a 50% increase of the concentration of heavy-isotope sulfur.

Some investigators believe [3] that high $\delta^{34}\text{S}$ values in considerable amounts of sulfides indicate late diagenetic or epigenetic origins and are associated with the activity of sulfate brines typical for oil-bearing provinces. Brines associated with oil deposits are of two types: mother liquor or sedimentary brine, and lixiviation brine. Hot solutions of this kind have been described in the Neogene of California and, in the Soviet Union, in east Turkmenia, the Apsheron oil and gas province, Cheleken Peninsula [4], and other areas [13]. In particular, the hot brines of Cheleken are enriched in sulfate with an isotopic composition including extremely heavy (up to +30%) sulfur. The hot epigenetic and circulating solutions enriched in heavy sulfur with sulfate contents of several grams per liter can be a source of a considerable volume of sulfide mineralization, because the amount of circulating water is incomparable to the volume of interstitial water. When these waters come in contact with organic matter, the redox environment is changed sharply and a rapid almost complete sulfate reduction takes place. The resulting reduced forms of sulfur have a heavy isotopic composition ($\delta^{34}\text{S}$ can be as high as 20% and much more).

Objections have been raised to the hypothesis of the participation of brines in ore mineralization. First, certain structural and morphogenetic features (the widespread finely scattered pyrite insets, nodular forms, relict primary-sedimentary structures and texture, such as collomorphous, framboidal, etc.) rule out an epigenetic character of sulfide formation.

Secondly, for sulfur compounds formed from sulfatic epigenetic brines one should expect either a homogeneous isotopic composition or changes following some regularity (for example, down the section, laterally, or according to structural pyrite varieties) corresponding to different stages and periods of hydrothermal processing. However, we observe a here a considerable scatter and "randomness" of findings. Such a picture is more likely to reflect a low-temperature biogenic sulfate reduction.

Some investigators believe [16] that sulfides with a heavier isotopic composition could be formed by bacterial sulfate reduction if its rate is relatively slow. As has been established [3, 9, 18-20], the mechanism of isotopic separation in sulfate reduction is based on a kinetic effect. The kinetic isotopic effect is caused by the different rates of chemical reactions for light and heavy isotopes [11]. It is particularly obvious in low-temperature sulfate reduction. Accordingly, the more intensive the process, the faster the bacterial sulfate reduction and the less significant the isotopic separation. Theoretically, at certain rates (or intensities) of bacterial sulfate reduction no separation should occur. The intensity of sulfate reduction is apparently affected by a large number of factors. Unfortunately, there have been few studies of these factors, and generally it is premature to discuss these hypotheses.

Concerning the hypothesis of the participation of mineralized brines in sulfide mineralization, we should note that there are ways to correlate isotopic-geochemical and lithologic data. It is possible, for example, that brines flowed into unconsolidated high-organic sediment (such as through discharge of Cheleken-like brines on the Caspian seafloor, or the actual

situation in sediments of the Gulf of California). Brines with sharply different salinities do not mix [2]. As a result, discharge of highly mineralized brines into a relatively low-salt basin should create a peculiar environment for diagenesis. The main condition for this hypothesis is the existence during the early Precambrian of highly mineralized isotopically heavy brines, that is, evaporitic and protoevaporitic sediments,

Distinctions between sedimentogenetic and diagenetic or epigenetic processes cannot always be drawn reliably even in relatively young formations, let alone early Precambrian rocks.

* * *

Certain conclusions can be drawn from this study.

1. The influence of regional metamorphism on alteration of primary isotopic ratios is insignificant; for this region it is possible to reconstruct processes of the premetamorphic stage of rock formation.

2. Sulfur from the sedimentary cycle participated in the development of sulfide mineralization; no effect of sulfur from the earth's interior has been established.

3. Igneous rocks (gabbro-diabases) contaminate sulfur from the surrounding strata.

LITERATURE CITED

1. P. A. Borisov, Karelian Shungites [in Russian], Gosizdat, KFSSR, Petrozavodsk (1956).
2. M. G. Valyashko, "Genesis of brines of the sedimentary shell," in: Proceedings of a Geochemical Conference on the Chemistry of the Earth's Crust, Dedicated to the 100th Anniversary of V. I. Vernadskiy, Vol. 1 [in Russian], pp. 253-277 (1963).
3. V. I. Vinogradov, "The role of the sedimentary cycle in the geochemistry of sulfur isotopes," Tr. GIN Akad. Nauk SSSR, Issue 351 (1980).
4. V. I. Vinogradov, I. V. Chernyshev, and L. L. Shanin, "Isotopic composition of sulfur in recent metalliferous hot springs on Cheleken," Geol. Rudn. Mestorozhd., No. 3, 84-88 (1969).
5. L. L. Galdobina, "Metallogeny of shungite-bearing and shungitic rocks of the Onega trough," in: Materials on the Metallogeny of Karelia [in Russian], Karel'sk. Fil. Akad. Nauk SSSR, Petrozavodsk, pp. 100-113 (1982).
6. L. L. Galdobina and A. I. Golubev, "Carbonaceous (shungite-bearing) rocks of the Onega trough and their metallogenic specialization," in: The Metallogeny of Karelia [in Russian], Karel'sk. Fil. Akad. Nauk SSSR, Petrozavodsk, pp. 133-143 (1982).
7. L. L. Galdobina, Yu. K. Kalinin, S. V. Kupryakov, and P. F. Ivankin, "Endogenous origin of shungitic rocks of the Proterozoic in Karelia," in: Second National Conference on Carbon Geochemistry [in Russian], GEOKhI, Moscow, pp. 79-81 (1986).
8. The Geology of Shungite-Bearing and Volcanic Sedimentary Formations of the Proterozoic in Karelia [in Russian], Karelia, Petrozavodsk (1982).
9. Geochemistry of the Sedimentary Process in the Baltic Sea [in Russian], Nauka, Moscow (1986).
10. A. I. Golubev, A. M. Akhmedov, and L. L. Galdobina, Geochemistry of Black Schist Complexes of the Lower Proterozoic in the Karelia-Kola Region [in Russian], Nauka, Leningrad (1984).
11. V. A. Grinenko, and L. N. Grinenko, Geochemistry of Sulfur Isotopes [in Russian], Nauka, Moscow (1974).
12. A. A. Migdisov, S. L. Cherkovskii, and V. A. Grinenko, "Isotopic composition of sulfur in humid sediments correlation with their formation environment," Geokhimiya, No. 10, 1482-1503 (1974).
13. R. G. Pankina, Geochemistry of Sulfur Isotopes of Oil and Organic Matter [in Russian], Nedra, Moscow (1978).
14. V. A. Sokolov, L. L. Galdobina, and V. I. Gorlov, "Geology of the Upper Yatulian-Suisar-ian," in: Problems of the Geology of the Middle Proterozoic in Karelia [in Russian], Karel'sk. Fil., Akad. Nauk SSSR, Petrozavodsk, pp. 133-144 (1972).
15. S. O. Firsova, G. V. Shatskii, and S. V. Kupryakov, "On shungitic breccias," in: Second National Conference on Carbon Geochemistry [in Russian], GEOKhI, Moscow, pp. 285-287 (1986).
16. F. V. Churkhrov and L. P. Ermilova, "New data on the genetic significance of the isotopic composition of sulfur in nodules," Litol. Polezn. Iskop., No. 3, 76-84 (1973).
17. Shungites: A New Source of Carbon [in Russian], Kareliya, Petrozavodsk (1984).

18. A. G. Harrison and H. G. Jhode, "Mechanism of the bacterial reduction of sulfate from isotope fractionation studies," *Trans. Faraday Soc.*, 54, No. 421, 84-92 (1958).
19. J. W. Hunt and J. W. Smith, "Sulfur $^{34}\text{S}/^{32}\text{S}$ ratio of low-sulfur Permian Australian was in relation to depositional environment," *Chem. Geol. Isotope Geosci. Section*, 58, No. 1/2, 137-144 (1985).
20. H. R. Krouse, R. G. McCready, S. A. Hussain, and J. N. Campbell, "Sulfur isotope by salmonella species," *Canadian J. Microbiology*, 13, 21-25 (1967).
21. H. G. Thode, "Sulfur isotope fractionation in nature and geological and biological time scales," 3, No. 5, 235-243 (1953).

ANCIENT ARID ELUVIUM OF TIANSHAN

V. S. Erofeev and Yu. G. Tsekhovskii

UDC 551.311.233(575)

This article first considers in detail questions on the structure, composition, mineral transformations, and formation conditions of late Cretaceous-Eocene and Miocene arid weathering crusts and soils in the Tianshan territory. Different groups and types of ancient arid eluvium are recognized. It is shown that the processes of ancient arid weathering led not only to the physical denudation of the source rocks, but also often accompanied their greater deep chemical transformations (including hypergene metasomatism, clay formation, etc.).

In theories of lithogenesis the concept predominates that weathering processes in arid landscapes proceed weakly, and that they are principally limited by physical forms of transformation of the aluminosilicate substrate [6, 15, 17, 26, 31, 32]. Therefore, mainly gruss eluvium forms here on the crystalline basement rocks, and rock crushing, migrations of carbonates, sulfates, and chlorides in the profile, and their metasomatic displacement by aluminosilicates are considered the basic processes of arid weathering. There is a point of view [17, 26] that clay formation by processes of arid weathering of aluminosilicate rocks does not occur at all, and the presence of clay material in arid soils is connected with its inheritance from the maternal substrate [8, 9]. However, the majority of investigators consider that arid weathering of rocks accompanies their deeper chemical transformation: hydrations, partial leaching and formation of hydromica-montmorillonitic clay minerals [7, 15, 31, 32] or sometimes palygorskites [2, 18]. A different scale of processes for arid clay formation is proposed: from extremely limited, with slight inclusion of newly formed clays in eluvium [32], to significant, with appearance of arid clay hydromica-smectite weathering crusts [7, 11, 12, 15, 21, 22, 31]. Moreover, arid eluvium (compared to humid) is still insufficiently studied and characterized in the geological literature. At this time, knowledge of arid weathering processes makes it possible to resolve a series of keenly discussed questions on the genesis of the arid sequences correlated to them. For example, in basins of Tianshan, the thickness of arid continental sedimentary deposits of upper Cretaceous, Paleogene and Neogene age reaches 1000 m [12, 28, 34, 36]. Sequences of clays, marls, limestones, gypsums, and salts often occur in their composition, in addition to coarsely detrital rocks. Analyzing the ancient sources of drift contributing to their formation, all geologists working in Tianshan acknowledge the important role of processes of physicochemical weathering in the limits of the ancient arid land. However, their intensities are differently judged.

Some authors [35] consider that mainly limestones, dolomites, and pyritized clay shales underwent chemical weathering in regions of denudation in the middle of crystalline basement rocks. This established the alkaline and alkaline-earth metals and sulfate-ions necessary for the subsequent thickening in the sedimentation basins of carbonate or sulfate salts. At the same time, the fine grained products of physical weathering of rocks with their subsequent

Geological Institute, Academy of Sciences of the USSR, Moscow. Altai Department of the Institute of Geological Sciences. Academy of Sciences of the Kazakh SSR, Ust'-Kamenogorsk. Translated from *Litologiya i Poleznye Iskopaemye*, No. 1, pp. 29-48, 1990. Original article submitted December 21, 1988.

sedimentational-diagenetic reworking into clays entered the sedimentation basin from drift sources.

Other geologists [11, 12], acknowledging the authigenic nature of the sepiolite-palygorskite clays gathering in the accumulation basins, at the same time consider that the basic mass of pelitic rocks (montmorillonite-hydromica composition) is allothigenic, and that it was introduced from ancient weathering crusts.

We carried out determination of the above noted disputed questions on arid weathering on an example studying ancient Mesozoic-Cenozoic weathering crusts of Tianshan.

CHARACTER OF DISTRIBUTION OF UPPER MESOZOIC-CENOZOIC ARID WEATHERING CRUSTS OF TIANSHAN

It is possible to unify the Mesozoic-Cenozoic weathering crusts of Tianshan known at the present time into two large groups: the humid, lower Mesozoic, and the arid, upper Mesozoic-Cenozoic.

The lower Mesozoic humid kaolinitic and locally lateritic weathering crusts of Tianshan often are detected here at the base of Triassic-Jurassic terrigenous-carbonaceous sequences, which protected them from erosion in subsequent tectonically active epochs. These weathering crusts are characterized in publications of V. I. Troitskii [27], B. A. Bogatyrev [3], and others. This eluvium is developed principally in the Jurassic basins of southwest Tianshan.

Replacement of an earlier existing humid climate by an arid climate generally occurred in the Tianshan territory in the late Jurassic [23]. Farther along in the Cretaceous, Paleogene, and Neogene, an arid climatic environment was established in the characterized region (with variations of climate from extra-arid to moderately arid), and typically arid deposits were accumulated (continental reddish-carbonate molasse, sometimes with inclusions of lacustrine gypsums, salts, or marine terrigenous-carbonate deposits, often with sequences of dolomites, gypsums, and salts. Monographs of V. A. Babadagla and A. Dzhumagulov, A. G. Babaev, N. N. Verzilin, M. R. Dzhalilov, L. B. Rukhin, A. V. Sochava, and many others are devoted to the complete description of Cretaceous arid deposits in the characterized region and their formation conditions. Similar publications are also available for the Cenozoic history of the given region [4, 19, 28, 35]. The evolution of the Cretaceous and Paleogene-Neogene arid paleo-landscapes of Tianshan and the surrounding regions is shown in the atlas of maps [1], and paleo-climate reconstruction was conducted by V. M. Sinitsyn [23].

There are no reliable data in the literature on the epochs of existence of a humid climate in the Cretaceous-Paleogene-Neogene development history of Tianshan. It is true, the Yakkakhan bauxite ore shown is noted in the southwest in one of the parts of the Zeravshansk range under Cretaceous deposits on Devonian rocks. These are conditionally dated as Aptian-Albian [3]. However these bauxites are apparently Triassic-Jurassic, since the stratified lower Cretaceous deposits of southwest Tianshan are generally arid formations, which precludes finding bauxites in them.

Thus, within the limits of ancient Tianshan, different variations of arid weathering and lithogenic processes predominated throughout the Cretaceous, Paleogene, and Neogene epochs. This has favored their reconstruction and knowledge.

Upon Cretaceous-lower Paleogene and lower Neogene arid weathering crusts are most completely and variously represented in the limits of Northern Tianshan. The upper Cretaceous-lower Paleogene deposits here crown remnants of a widely developed weathering surface which is pre-upper Cretaceous-lower middle Paleogene. They are developed on metamorphic sedimentary or magmatic rocks of the Paleozoic basement. Moreover, they form horizons of buried eluvium in deposits of upper Cretaceous and Paleocene-Eocene suites of the characterized region. The lower Neogene deposits represent horizons of buried eluvium in sections of Miocene deposits.

The first indications of the presence of ancient (pre-Tertiary) weathering crusts in Tianshan were published long ago [19, 34]. Later, V. N. Shcherbina [35] briefly characterized ancient eluvium here in the process of investigating Neogene salt-bearing sequences. However, eluvium refers to both strictly weathering crusts, and their redeposited products: clays, sands, marls, etc. Therefore this geologist estimated the thickness of such an eluvial-sedimentary body in the hundreds of meters. Subsequent investigations [28, 36, etc.] con-

the presence of weathering crusts on basement rocks at the base of the Paleogene-Neogene continental series of Tianshan (the Kirgiz Red-bed Complex) and their correlated erosion products, which are distinguished in the composition of the upper Cretaceous-Eocene Sulutereksk or Kokturpaksk suites. At the same time, we note that up to this time there are practically no publications concerning the detailed structure and mineral composition of the ancient alluvium of Tianshan, and its specifics caused by existence of arid paleoclimate are not generally indicated. Only A. V. Sochava [24, 25] noted that arid carbonate crusts are present here, and briefly characterized them in a series of regions, subdividing them into interformational (developed on rocks of the Paleozoic basement) and intraformational (developed on Cretaceous argillites). In addition, there is brief information available on arid carbonate or gypsum intraformational crusts of Tianshan in the Miocene Irtashsk suite [5], as well as in the late Cretaceous to Eocene deposits [11]. In the process of investigations in Tianshan territory, we studied both of the above noted time stages of intensive formation of ancient arid alluvium: 1) Miocene, 2) late Cretaceous-Eocene. The first reflected an accumulation of Miocene deposits that are correlated to the given alluvium of the Irtashsk and Serafimovsk suites. It was established that weathering crusts of both of the above named stages formed not only on Paleozoic basement rocks, but also on Miocene or late Cretaceous-Eocene deposits (in the latter case they represent intraformational deposits in relation to the above named formational suites). However, primary attention was concentrated on study of upper Cretaceous-Eocene weathering crusts, which are widely and variously represented, and which are characterized by the great intensity of eluvial processes. Sections of the given weathering crusts, connected with pre-Tertiary surfaces from the weathering of the folded Paleozoic basement, have been noted here by many geologists [19, 28, 34, 36]. The more intense tectonic motion of the Cenozoic, which created the contemporary aspect of the mountain country, led to arching and arch-block deformations of the ancient peneplain. Those parts which entered into regions of arch-block uplifts were subjected to the action of energetic erosional-denudation processes: the deep linear separation of ancient weathering surfaces into blocks, and denudation with their loose products of weathering crust. Therefore only scarce remnants of the lower Shebnisto-Dresvyanoi eluvial zones are observed on uplifted relicts of ancient weathering surfaces in a majority of cases. In regions where the weathering surface was buried under a Cenozoic sequence, the thickness of the eluvium often increases and its upper zones are apparent: clay, calcareous, gypsum. Therefore relatively good sections of ancient alluvium are observed mainly on the periphery of intermountain basins, where the bases of Cenozoic deposits are also often exposed. However, a test also shows that here the weathering profile is generally better preserved and represented in places where the crust is principally covered by clay deposits. This is clearly expressed under clay sequences at the tops of late Cretaceous-lower Paleogene. In such regions where coarsely detrital rocks of a higher stratigraphic stage of Cenozoic cover are deposited onto the Paleozoic basement, preservation of the ancient crust deteriorates (up to complete erosion).

At the present time, sections of weathering crust are established with varying degrees of completeness in the near-flank parts of almost all large intermountain basins (Issyk-Kul'sk, Vostochno-Chuisk, Narynsk, etc.) and on ridge slopes adjacent to them. This fact, with consideration of the finding of "roots" of structural alluvium on denuded relicts of the peneplain leads to the conclusion that the ancient weathering crust of Tianshan had at some time a wide, probably areal, character. Its study made it possible to distinguish a series of characteristic varieties, which, in agreement with the classification of I. I. Ginzburg [7], must be related to infiltrated weathering crusts. In this article we apply the more divided scheme of alluvium subdivision of B. M. Mikhailov and B. V. Kulikova [17]. We include the characterized products of ancient hypergenesis in two genetic groups: 1) arid weathering crusts of protrusions of crystalline basement rocks; 2) arid weathering crusts on the sedimentary cover (in the characterized region they are intraformational).

ARID WEATHERING CRUSTS OF PROJECTIONS OF CRYSTALLINE BASEMENT ROCKS

There are two basic types of weathering crusts apparent in the given group: a) clay gypsum-calcareous; b) clay-gruss, armored by a calcareous plating.

Clay Gypsum-Calcareous Weathering Crust. The given type of ancient weathering crust is widely developed on the peneplains of Tianshan. However, a complete profile of this crust does not often occur. It is possible to observe a similar section of alluvium, for example, in the high mountain basin of Kokdzharsu (Northern Tianshan) in outcrops of the left flank of the Teepshi river valley (Fig. 1). At 1.5-2 km above the river mouth, its valley is cuts an arching uplift, the core of which reveals an ancient peneplain surface made up of

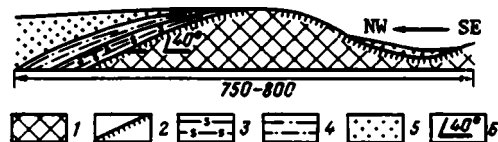


Fig. 1. Section of the left flank of the Teepshi river valley and 1.5 km above the mouth (Kokdzharsu basin, Northern Tianshan). 1) Carboniferous arkose sandstones; 2) weathering surface recorded by red-mottled weathering crust; 3-4) lower Paleogene deposits; (3) packets of red clay gypsums; 4) packets of dark-red siltstones; 5) upper Oligocene red-bed sandstones; 6) slope angle of the ancient weathering surface.

dislocated Paleozoic rocks and crowing weathering crusts. On its limbs this weathering surface is buried under layers of Paleogene deposits. Ancient weathering crusts are well exposed on the northwest slope of the uplift, immediately under deposits of gypsum-bearing clay redbeds of the lower Paleogene.

Grey colored arkose sandstones of the lower Carboniferous (varied grained, from fine to coarse, principally feldspar-quartz composition) serves as the maternal substrate of the eluvial profile (Fig. 2A). The detrital material has a varied degree of rounding and poor sorting. Quartz (up to 40%) and feldspars (up to 30%) predominate in its composition. Fragments of shales, sandstones, and granites seldom occur. The cement is of silt-clay, dense, and contact up to pore. Fine mica leaves are noted.* A profile of the weathering crust formed on sandstones has a thickness of 8-12 m and is subdivided into three zones (from the bottom upwards).

Zone I is the lithic eluvium. In its limits sandstones of an off-white color predominate, and there is a thickly dispersed network of veins of dense pelitomorph limestone of white and pale yellow color (thickness from 1 to 15 cm). The most coarse and thick carbonate veins are oriented in a vertical and steeply sloped position, and they are united among themselves by a system of subhorizontal and slightly sloped thin veins and stringers. The coarsest of them sometimes cut into the non-weathered sandstones up to a depth of 10 m. Pelitization of the feldspar grains is noted, and there is corrosion and cracking of the quartz fragments, which are generally of light color. The degree of alteration in the maternal rock increases from below upwards in the profile. The sandstones become weakly lithified (disintegrating in the hand). Yellowish-brown spots of iron oxides are apparent in them. The network of carbonate veins gradually becomes more frequent and possesses the form of stockwork; irregular spots of pelitomorph calcite are apparent. In the mass of pelitomorph carbonate, sandstone relicts half-replaced by the carbonate-clay mass are frequent. There are sections where relatively fresh rocks are preserved. Table 1 shows the results of silicate analysis of rocks from the characterized zone (samples Tsh-76-14a, Tsh-76-14; Tsh-76-16, see Fig. 2, A). The thickness of the described zone reaches 5.0-6.5 m.

Zone II is the disintegrated clay-gruss products of sandstone weathering. This zone is connected with the lower lying zone by a gradual transition. The rocks are divided into yellow and yellow-brown color. Textural features of the maternal rocks are still preserved in individual parts of the zone's bases, since the source rocks are transformed into gruss between them and higher in the profile. Feldspar grains are almost completely pelitized, but preserve their form; quartz fragments are strongly fractured and corroded by the carbonate-clay mass. Upwards in the profile, ferruginization spots become more frequent and darkly colored. Clay is apparent in the cement of the gruss, sometimes with fine gypsum crystals. Strongly weathered relict sandstone fragments are locally preserved. They are easily mashed in the hand. The veins of pelitomorph calcite characteristic of zone I disappear in the gruss products of weathering. Instead of them, the pelitomorph calcite is isolated in the form of scarce fine disseminations in gruss or (in the upper zone) forms nodules. The pelitic fraction of zone II rocks, judging by diffractograms (see Fig. 2B, sample Tsh-19), is made of hydromica-smectite clay material (probably in the form of a mixed layered hydromica-montmorillonite mineral with a high content of the smectite components). Quartz and calcite are present as admixtures.

*Study of rock under microscope was conducted by G. S. Karlova.

Results of silicate analysis of rock samples from Zone II (see Table 1, samples Tsh-76-17, -18, -19, -20) indicate that in comparison with samples from Zone I, there is a fall in the content of SiO_2 , K_2O , and Na_2O , and an increase in the value of Al_2O_3 , Fe_2O_3 , and TiO_2 . This may indicate the start of the process of chemical weathering and argillization of the source rock. The increase in the value of CaO , SO_3 and p.p.p. in the series of samples in the given zone indicate the presence of an admixture of calcite or gypsum. The thickness of the described zone reaches 1-1.5 m.

Zone III are the clay products of weathering. It crowns the alluvial profile, connected by a gradual transition from lower lying rocks and made up of red-brown sandy clays with spots and subvertical bands of green-grey color. Textural-structural signs of the maternal rocks are generally lost. Feldspars and micas are transformed into clay. Diffractograms of the pelitic fraction of zone III rocks have a unique aspect similar to sample Tsh-76-23 (see Fig. 2B). In them, the smectite component generally clearly dominates the hydromica component (probably in the composition of a mixer layer hydromica-smectite mineral). Smectite reflections in the low angle region are represented in the form of wide peaks with formation of a series of blending packets. This may indicate both the low degree of its crystallization, and the possibility of input of a mixed layered hydromica-smectite mineral into the composition. Thermograms of the clay fraction are also unique and resemble sample Tsh-76-23 (see Fig. 2B). This underscores their principally smectite composition.

From the results of silicate analysis (see Table 1, samples Tsh-76-21, -22, -24) it is apparent that in zone III, in comparison with zone II, the decrease in SiO_2 and K_2O contents, and the growth of Al_2O_3 and Fe_2O_3 values continues. This is connected with the continuing argillization of the source rocks and the formation of smectites. Simultaneously some growth in the values of MgO , SO_3 , and sometimes CaO , MgO , and Na_2O is recorded. This may indicate salinization of rocks in the profile with accumulation of carbonates or gypsum. The thickness of this zone is 2-3.5 m.

On the whole, the characterized profile records from below upwards the weathering and argillization of the source sandstones. This is accompanied by the preservation of quartz grains, as well as by the carbonatization, gypsumization, or salinization of the rocks.

A 30 meter thick clay-gypsum packet of the lower Paleogene lies on the described profile of the weathering crust in the exposures on the left flank of the Teepshi river.

On the northern slopes of the Kirgzi range (Northern Tianshan) an ancient calcareous argillaceous weathering crust analogous to the above described profile was formed on lower Paleozoic gneisses. It was studied in exposures of left flank of the Shamsi river valley at 1 km below the mine of the same name (Fig. 3a). Here bleached and ferruginized gneisses already touched by weathering are detected in the cuttings. They are split by a thick network of carbonate veins with subvertical orientation. Higher, the bleached gneisses transform into a zone of clay-gruss-detrital eluvium with spots of ferruginization 1-1.2 m thick. A network of pelitomorphic calcite veins grows in it, which abruptly disappears higher in the profile at the transition to zone III. The clay rocks of zone III are enriched by quartz remnants and have a green-grey color (with brown spots). Fine calcite disseminations are often apparent in them, and calcareous concretions and gypsum layers occur. Here, apparently, only the lower part of the clay weathering products zone is developed, with a thickness of 0.8-1 m, while its tops are eroded. The erosion surface is clearly recorded by Paleogene sandstones with dense carbonate cement which lie on the crust.

The character of the mineral transformations and rock composition of the eluvial profile at the Shamsi mine was similar to those from the Teepshi river which are characterized above. In particular, results of silicate analysis of the rocks from the Shamsi profile (see Table 1) and thermograms of the clay fractions (see Table 1) and thermograms of the clay fractions (see Fig. 3, samples Tsh-36, -37, -38) indicate the presence of a substantially smectitic composition in the clay products of gneiss weathering. A general carbonatization of profile tops is noted (the CaO content locally reaches 14 and 17.9% here).

In Southern Priissyskul'sk it is possible to observe profiles of weathering crusts of similar structure (and their denudational relicts) developed on the boundaries. They are found on the peneplains of the Terskei-Altai range in the basins of the Tamga, Tossor, and Dzhetuyoguz rivers in the Boomsk ravine. In Central Tianshan on northern slopes of the Karatau range they armor the surface of the ancient peneplain and are well exposed here. Similar sections of red-bed clay gypsum-limestone weathering crust are also detected in many other regions of Tianshan.

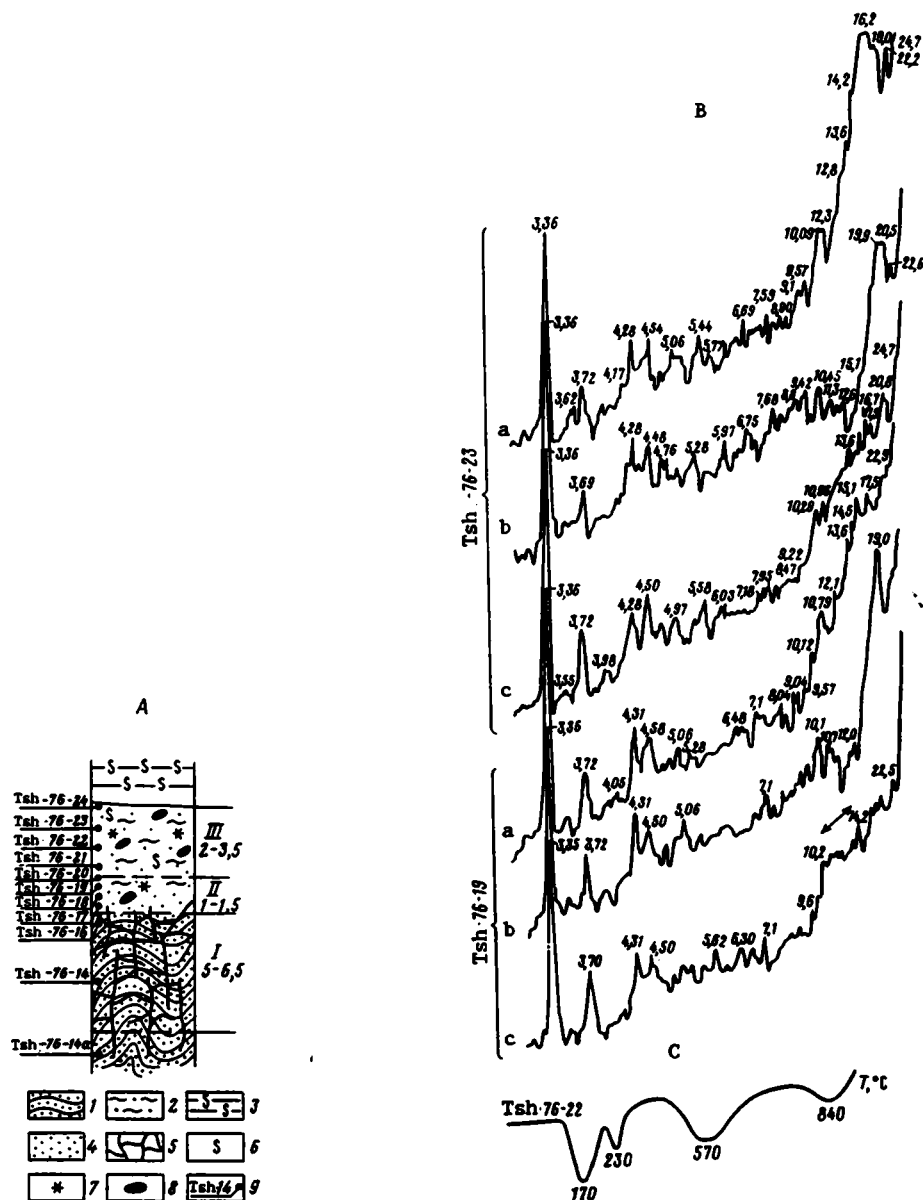


Fig. 2. Structure and composition of a clay gypsum-calcareous weathering crust in exposures on the left bank of the Teepshi river. A) section of the profile: 1) lower Carboniferous arkosiv sandstones; 2) eluvial clays; 3) clay gypsums; 4) eluvial grusses; 5) veins of pelitomorphic calcite; 6) gypsum inclusions and veinlets; 7) spots and bands of ferruginization; 8) calcareous concretions; 9) sample number and selection site; on the right are cited number of the material-genetic zone of the weathering crust (Roman numbers) and its thickness, m. B) diffractograms of clay fractions (a) natural samples; b) saturated by glycerine; c) heated to 350°). C) thermogram of the clay fraction.

Summarizing the concepts of arid weathering processes in clay calcareous-gypsum weathering crusts, we note that weathering of sedimentary and magmatic silicate rocks of the basement is accompanied both by physical and chemical denudation. Feldspars, micas, and dark minerals underwent destruction and argillization, and hydromica and smectites, as well as detrital quartz, accumulated in their weathering products.

Argillization of silicate rocks is accompanied by decrease in K_2O , MgO , and SiO_2 contents, and simultaneously by increase in the values of Al_2O_3 , Fe_2O_3 , and TiO_2 .

TABLE 1. Chemical Composition of Rocks of Clay Gypsum-Calcareous Weathering Crusts, Mass %

Samp. number	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	CaO	MgO	TiO ₂	MnO	P ₂ O ₅	K ₂ O	Na ₂ O	SO ₃	p.p.p.	Σ
Teepshi profile (see Fig. 2)														
Tsh-76-14 ^a	75,88	11,26	2,50	0,07	1,91	Not det.	0,08	0,04	0,03	3,74	1,72	0,19	2,94	100,55
Tsh-76-14	79,06	11,08	2,30	0,07	1,01	"	0,09	0,02	0,01	3,26	2,04	0,30	1,26	100,80
Tsh-76-16	64,50	10,21	1,50	0,07	10,58	"	0,10	0,04	0,02	2,40	0,15	0,36	9,16	99,45
Tsh-76-17	69,70	12,52	2,80	0,07	4,51	"	0,15	0,04	0,02	2,69	1,88	0,56	4,78	100,28
Tsh-76-18	72,06	15,02	2,50	0,07	0,67	"	0,25	0,05	0,01	2,35	1,30	0,72	3,94	99,66
Tsh-76-19	62,32	14,10	4,40	0,07	1,80	2,27	0,53	0,07	0,04	1,10	1,29	1,73	8,22	99,67
Tsh-76-20	62,56	13,46	5,05	0,07	2,25	2,43	0,56	0,05	0,04	0,62	1,56	1,20	9,00	100,05
Tsh-76-21	58,96	14,36	5,13	Not det.	2,03	4,05	0,63	0,16	0,05	0,43	1,35	1,56	9,84	100,11
Tsh-76-22	56,22	14,11	5,70	0,07	4,06	3,24	0,53	0,08	0,06	0,67	1,40	2,14	9,76	100,18
Tsh-76-23	60,02	15,00	5,05	0,07	1,58	3,08	0,63	0,13	0,04	0,38	1,88	1,90	8,88	100,54
Tsh-76-24	59,72	15,00	5,02	0,10	1,01	3,56	0,63	0,13	0,05	0,53	1,62	1,26	10,20	100,09
Shamshi profile (see Fig. 3, a)														
Tsh 34	66,00	14,50	4,64	0,79	1,55	2,90	0,81	0,09	0,09	2,69	1,45	Not det.	4,92	100,43
Tsh-35	65,52	12,67	3,21	2,37	1,85	4,09	0,85	0,06	0,12	2,59	1,83	"	4,22	99,38
Tsh-36	55,80	15,06	7,10	0,07	2,53	3,09	1,42	0,15	0,07	0,67	Not det.	"	14,20	100,16
Tsh-37	43,00	12,19	7,57	0,07	14,00	2,00	1,15	0,14	0,07	0,43	"	"	19,84	100,46
Tsh-38	42,22	9,08	4,60	0,07	17,90	2,70	0,50	0,24	0,04	1,34	"	"	20,84	99,53

Note. The analysis was conducted in the chemical laboratory of the Altai department of the K. I. Satpaeva IGN of the Academy of Sciences of the Kazakh SSR.

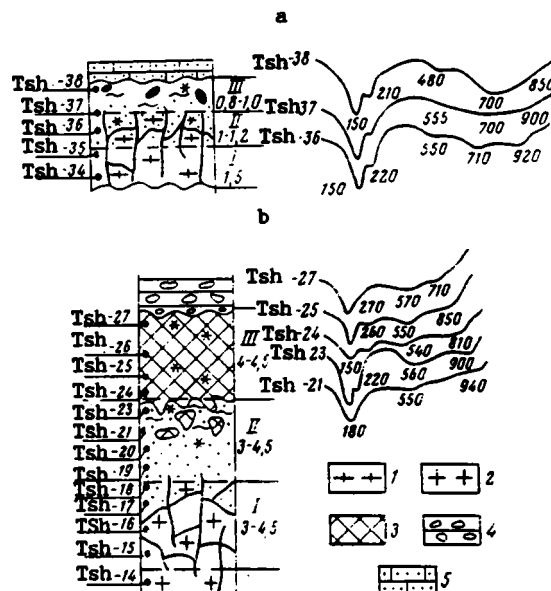


Fig. 3. Structure and composition of weathering crust of Northern Tianshan. a) profile of the calcareous-clay weathering crust on the left bank of the Shamsi river valley in the Chuisk basin; b) profile of a clay-gruss crust armored by calcareous plating in an exposure of the Aksaya and Aktereka interfluvium (Issyk-Kulsk basin). On the right, clay thermograms are shown. 1) lower Paleozoic gneisses; 2) granites; 3) calcareous plating; 4) Neogene conglomerates; 5) lower Paleogene sandstones with calcareous cement. For the remaining symbols see Fig. 2.

The behavior of iron compounds is significant. Releasing from the mineral lattice in conditions of alkaline arid weathering, it is stripped by migration mobility, rapidly transforms from ferrous to ferric forms, covering sand-clay weathering products by the thinnest bright red plating and gradually accumulating on them. The bright red color of the characterized ancient arid weathering crusts are connected precisely with this. Their erosion also led to the origin of the correlated red-bed deposits of the Sulutereksk suite in accumulative arid landscapes. It is also important to note that in the characterized arid eluvium (compared with humid) the scale of removal of alkaline and alkaline-earth metals released by silicate argillization decreases. The tendency towards a decrease of K_2O in the profile is most distinctly expressed, while the values of CaO , MgO , and Na_2O vary strongly, and often increase. This indicates carbonatization and salinization of eluvial rocks. This is not only connected with accumulation in the rocks of the above noted metals released by silicate weathering, but also primarily with their infiltrations into the eluviums by vertical capillary ascent and by the evaporation of waters due to lateral subflow with ground water in the hydromorphic stage of landscape development.

It was noted earlier that all of the ancient crusts we have studied were buried under sediments after their formation. These sediments guarded them from erosion. Parts of the uplifted regions where the initial automorphic weathering crusts formed began to experience lowering, and through this the automorphic weathering environment was changed to hydromorphic. Just in the concluding, hydromorphic stage of their formation (preceding the burial of eluvium by the sediments of the Sulutereksk suite) infiltration of components into the profile rapidly increased, and the most intense carbonatization, gypsification, and salinization of eluvial rocks occurred. The processes of salt infiltrations into the eluvial profile probably could be partially prolonged, even after further lowering of the characterized parts and transformation of denudational landscapes to accumulative. For strongly gypsified deposits on the eluvium surface or carbonate sediments of the Sulutereksk suite (see Fig. 2, 3) further gypsification or carbonatization of the profile may occur, along with infiltration of salts from the top downwards. However, the scales of similar epigenetic transformations were not generally great. The point is that deposits of the upper Cretaceous-Eocene Sulutereksk suite are generally represented [11, 19, 28, 36] principally by clay montmorillonite sequences of rocks only locally

containing small lenses of sands, silts, and gravels. These deposits, with thicknesses of several hundreds of meters, are an excellent regional water resistor. Therefore, conditions for underground water circulation in the given deposits were extremely unfavorable, and consequently the scales of epigenetic addition of components into the profile (or, conversely, their removal) were not great. In a majority of cases, deposits of similar clay sequences on an ancient eluvial surface assisted its conservation and isolation from the active influence of underground water.

Clay-Gruss Weathering Crust, Armored by Calcareous Plating. These weathering crusts occur on selected limited parts of the ancient peneplain of Northern Tianshan. The most complete and representative section is observed, for example, in horst uplift exposures in the Aksaya and Aktereka interfluvium on the southern shore of Issyk-Kul'. Uplift from the southwest is sharply limited by a terrace along a fault, and an ancient weathering surface with a bright red weathering crust, smoothly subsiding to the northeast at an angle of 25-30° is recorded opposite its flank.

Coarse grained biotite-hornblende granite (up to 8% hornblende) serves here as the maternal substrate of the weathering crust (see Fig. 3b). The basic weathering profile is the stony eluvium of zone I. Its structure is little distinguished from the zone of clay crust corresponding to it that was earlier characterized. The maternal rocks are bleached, acquiring an off-white shade, and they are disturbed by a system of vertical and sloping fractures, often filled by veins of pelitomorphic calcite. The calcite veins often increase towards the tops of the zone, they often run together into extremely wide (to 1.2-1.5 m) vertical cavities, filled by gruss and cemented by pelitomorphic calcite. Individual carbonate veins extend deeply into granites (sometimes more than 10-12 m).

In the bleached granites of zone I, the feldspars (plagioclase and potassium feldspar) are already pelitized to a significant degree. Iron hydroxides developed from hornblende and biotite, and they form films in the rocks' microfractures. The thickness of the zone varies from 3 to 5 m (locally greater).

The characterized eluvial profile (in distinction from the clay weathering crust) has a thicker zone of gruss weathering products (zone II). At the bases of the zone, the grusses often preserve the structural-textural signs of granites. The largest part of feldspar grains is pelitized. The rocks are even more bleached, acquiring a light grey tone. Upwards in the profile of the zone, a great quantity of clay components is apparent in the gruss as a consequence of the decomposition of the overwhelming mass of feldspars and micas. Some of these grains still retain a rather fresh aspect and preserve their initial form. Simultaneously, rusty-brown and brown spots of limonitization are apparent in the rocks, and pelitomorphic calcite forms a fine dissemination of a locally dense cement of gruss grains. The thickness of this zone is 3-4.5 m.

Higher, the zone of clay grusses is gradually (over a distance of 0.5-0.8 m) succeeded by calcareous plating. In so doing, the content of disseminated pelitomorphic calcite in the clay grusses is increased. This forms an initially friable and subsequently dense clay-carbonate cement. Simultaneously, rock is colored a brownish color due to appearance of films of iron oxides on clay and sand particles.

The massive carbonate plating which is crowned by weathering crust represents an aggregate of angular corroded grains of quartz and pelitized feldspar, cemented by a dense clay-carbonate mass. Very often rocks of the plating are split by tube-like canals of subvertical form, filled by precipitations of coarsely crystalline calcite (in voids from plant roots or fractures). This leads to the appearance of a subvertical texture in the plating rocks, by which they are generally sharply distinguished from the background of the surrounding sedimentary rocks, which have a principally subhorizontal layered texture. Calcareous nodules are often abundantly developed in the plating rocks. They are represented by tubular calcite concretions (d 0.5-2.0 mm) around tree roots, with a principally subvertical orientation (sometimes root-like spaces are filled by calcareous-clay material with crystallized calcite, rarely chalcedony).

Concretionary structure often predominates in individual parts of the plating. The calcareous concretions (up to 2-3 cm in size), as well as the calcite pelitomorphic mass surrounding them, contain grains of quartz and pelitized feldspar. The clay-calcareous rocks of the plating are often pigmented by iron hydroxides, giving the rock a generally red-brown color. The thickness of the calcareous plating is 40-45 m.

Thin lenses (0.2-0.3 m) of horizontally layered pelitomorphic or gravel to fine pebble rocks are often apparent in the cover by tracing the calcareous plating in space. Rewashed plating rocks dominate the composition of the gravel-pebble rocks; the given redeposited products have often again undergone calcareous cementation. Sometimes lacustrine mollusk fauna also occur in them (for example in the Toruaigyr section).

The layering vanishes in a series of portions in these rocks; traces of ancient plants, similar to those earlier described are again apparent, indicating soil formation in the sediments after their deposition. It is possible to consider the lenses of these horizontally layered sedimentary rocks with buried soil horizons in the cover of the characterized carbonate plating as the basal horizon of the sedimentary Sulutereksk suite and its age analogs, excluding profiles of carbonate crusts from the structure of the zone III profile.

Judging by thermograms (see Fig. 3b, sample Tsh-21-23), the pelitic fraction of zone II rocks, represents hydromica-smectite minerals. In agreement with data of silicate analysis (Table 2), a general tendency is noted upwards in the profile (from zone I to zone II, sample Tsh-21-23) towards a decrease in SiO_2 , K_2O , and Na_2O contents, and increase in the values of Al_2O_3 , Fe_2O_3 , CaO , MgO , and TiO_2 . This indicates the direction of chemical alteration of the substrate, as well as in the above described argillaceous weathering crusts.

In zone III of the calcareous plating (see Table 2, sample Tsh-24-27) calcareous metasomatism of the initial silicates is accompanied by a decline in the profile of SiO_2 , Al_2O_3 , K_2O , and Na_2O values, and growth in values of CaO (up to 21.67%), as well as (to a lesser degree for a majority of the zone's samples) values of MgO , Fe_2O_3 , and TiO_2 . The composition of clay components, judging by thermograms (see Fig. 3, b) are hydromica-smectite.

Thus, in the process of formation of a grass-clay crust (with calcareous plating) on gneisses, calcitic metasomatism of source aluminosilicates occurs at the tops of the profiles, as well as their argillization with formation of hydromicas and smectites. The given type of ancient eluvium of Tianshan resembles deposits characterized by many geologists as calcareous or carbonge weathering crusts, caliche, and calcretes that are developed on crystalline rocks of the basement. Their mineral forming processes are completely considered in [10, 16, 17, 18]. However the calcareous weathering crusts of Tianshan are distinguished by the presence of a typically soil texture in the cover of the profiles and by clear manifestation of processes of aluminosilicate argillization in the middle parts.

Formation of thick carbonate weathering crusts is correctly connected with the lateral introduction of carbonate ground water into the profile under conditions of super-aquatic landscapes [10, 18]. V. V. Dobrovol'skii [10] distinguishes two stages in their formation. They are the initial stage: formation of a carbonate plate in the middle part of the eluvial profile; and the concluding stage; uplift of the territory, when the upper friable part of the eluvium was eroded, and thus, the plate was apparent on surface, armoring it.

We propose a somewhat different formation mechanism for the characterized calcareous weathering crust of Tianshan. In the first stage, probably only clay grasses formed, by automorphic weathering.* They were possibly weakly carbonate. In the following descriptions of the territory, automorphic weathering was replaced by hydromorphic, which sharply intensified the capillary and lateral flow of carbonates into the profile, assisting in the formation of calcareous plating in the cover of the eluvium. Under conditions of irregular burial of the basin floors of accumulative regions, along with general expansion of the eluvial profile downward, its increase upwards also sometimes occurred as a result of cyclic redeposition of the plating rocks (locally with supplemental input of sand-clay material externally) and their subsequent weathering.

The ancient hypogene formations considered in this correspond totally to the concept of infiltrated weathering crusts introduced long ago by I. I. Ginzburg [7]. The characterized eluvium was formed due to physical and chemical weathering of Paleozoic basement rocks on the surface of an upper Cretaceous peneplain in just pre-Tertiary time (before the deposition of the essentially clay sediments of Sulutereksk suite, which generally conserve the eluvial profile). Carbonates or sulfates simultaneously entering the profile (due to capillary or lateral flow) were involved in processes of eluvium formation, and they underscored its textural features (forming precipitations of subvertical form).

*Concepts on automorphic and hydromorphic environments are taken from study on the geochemistry of landscapes (data of M. A. Glazovskaya and A. I. Perel'man).

TABLE 2. Chemical Composition of Clay-Gruss Rocks Armored by Calcareous Plating of Weathering Crust, mass, %

Samp. number	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	CaO	MgO	TiO ₂	MnO	P ₂ O ₅	K ₂ O	Na ₂ O	p.p.p.	Σ
(Aksai-Akterk, see Fig. 3, b)													
Tsh-14	71,51	14,80	1,72	0,86	1,68	0,36	0,30	0,06	0,06	4,42	3,45	1,46	100,69
Tsh-15	70,64	14,95	2,12	0,50	1,80	0,50	0,22	0,06	0,05	4,80	3,29	1,74	100,61
Tsh-16	70,76	14,19	2,15	1,07	1,57	0,64	0,35	0,06	0,07	4,80	3,07	1,84	100,57
Tsh-17	72,12	14,04	2,04	1,72	1,46	None	0,27	0,06	0,05	4,60	3,07	1,96	100,39
Tsh-18	71,08	13,89	1,21	2,37	1,90	0,36	0,38	0,11	0,06	3,79	2,86	1,54	99,59
Tsh-19	69,56	15,10	2,94	0,36	1,35	0,90	0,35	0,04	0,05	4,32	3,07	2,58	100,62
Tsh-20	70,38	14,65	2,85	0,14	1,80	0,74	0,27	0,05	0,03	3,84	2,75	3,06	100,56
Tsh-21	66,86	16,31	2,60	0,36	2,31	0,74	0,35	0,05	0,03	3,72	2,86	4,26	100,45
Tsh-23	69,06	13,59	2,69	0,14	1,88	1,16	0,35	0,05	0,03	4,32	0,70	6,32	100,29
Tsh-24	44,14	10,12	2,60	0,07	19,71	0,36	0,35	0,06	0,05	2,16	0,97	18,66	99,25
Tsh-25	45,30	10,57	3,28	0,23	16,48	1,80	0,43	0,06	0,10	2,21	0,86	18,25	99,56
Tsh-26	51,80	12,55	4,27	0,36	10,00	2,50	0,50	0,07	0,09	2,59	None	14,56	99,27
Tsh-27	38,78	9,12	2,44	0,21	21,67	1,70	0,27	0,04	0,06	2,11	"	22,74	99,23

It was established that sediments of the Maastrichtian-Eocene Sulutereksk suite progressively lap onto an upper Cretaceous weathering surface by various stratigraphic horizons (up to the most youngest, which corresponds to the upper half of the Eocene). Therefore, the upper age boundary of weathering crust formation on a given leveling surface can vary accordingly and sometimes rose to the Paleocene or middle Eocene.

In the process of considering data on the characterized arid eluvium of Tianshan, a question arose that the considered clay weathering crusts could have initially been formed in humid landscapes, and only later, in immediately pre-Tertiary time, transformed their carbonatization or gypsification. The following data contradict this idea:

1. We noted above that in late Cretaceous time (when the pre-Tertiary leveling surface of Northern Tianshan originated with a cover of arid weathering crusts), and even long before (in the early Cretaceous), an arid type of paleo-climate stably predominated over the characterized territory. Therefore the montmorillonitic weathering crusts forming here encircle the synchronous typically arid deposits. The lower Mesozoic humid weathering crusts (lateritic or kaolinitic) widely developed in some regions of Tianshan (its southern part) crown the more ancient leveling surface and lie at the base of Triassic-Jurassic coal-bearing deposits. Consequently, the stratigraphic position and territorial correlation of the characterized arid and humid weathering crusts of Tianshan are completely different.

2. A series of special investigations [13, 14, 20, etc.] indicates that distinct signs which make it possible to identify the first stage of eluvium formation in humid landscapes are preserved in cases of superposition of subsequent arid weathering processes onto relict ancient humid eluvium. These are the presence of kaolinitic clays, replacement of iron-bearing and strongly iron-bearing parts with concretions of goethite or hematite, and other signs which are preserved in spite of later carbonatization or salinization of the rocks of the profile. The characterized clay arid eluvium has an entirely different aspect. Any relict traits of humid weathering are completely absent in them, and there are conversely many signs characteristic of a stable arid weathering character. This is confirmed not only by increased carbonatization or gypsification of the rocks of the profile, but also by the more uniform hydromica-montmorillonite composition of clays. The practically complete absence of migration of iron compounds in the profile is also an indicator. Separating out from the mineral lattice in the process of arid weathering under conditions of alkaline medium, they did not undergo migration and concentration, and they at once precipitated in form of fine films, evenly coloring the initial clay substrate into a bright cherry-red color.

INTRAFORMATIONAL ARID WEATHERING CRUSTS ON SEDIMENTARY COVER

This variety of eluvium generally records intra- or interformational interruptions of sediment accumulation. Separating similar types of eluvium forming on a friable sedimentary substrate, in humid landscapes, this has different names: Neo-eluvial (B. B. Polynov, V. P. Kazarinov), interformational and intraformational weathering crusts (I. I. Ginzburg, N. P. Kheraskov, V. N. Razumova, A. V. Sochava), weathering horizons (V. S. Erofeev, Yu. G. Tsekhovskii), fossil or buried soils (V. I. Chalyshev, A. P. Feofilova), weathering crusts on sedimentary cover (B. M. Mikhailov and G. V. Kulikova). Calcareous, gypsic, or salt-bearing

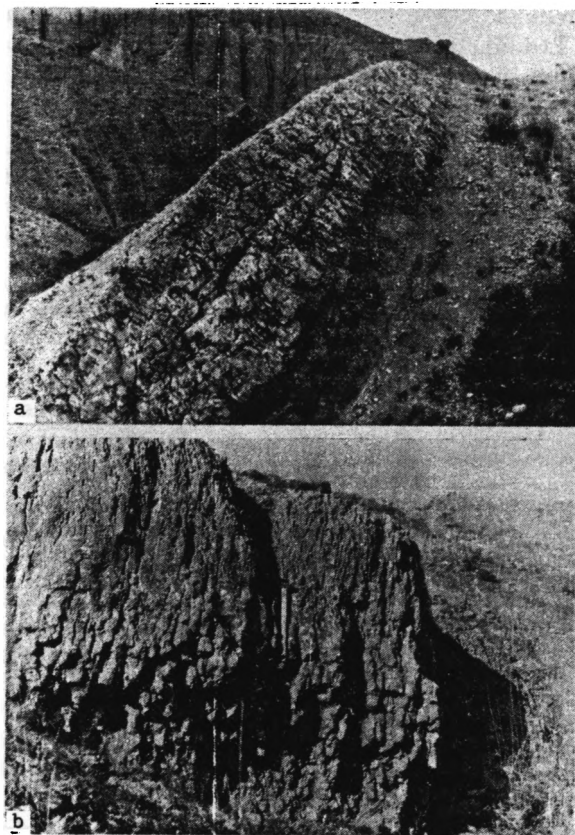


Fig. 4. Intraformational calcareous weathering crusts (caliche) of Tyan'-Shan, which form terraces in red-bed clays. a) in steeply dipping rocks of the late Cretaceous-Eocene Sulutereksk suite (the Chuisk basin); b) in the Miocene Irtashsk suite (Pretashkent basin at the Kysylsu river, right fork of the Chirchik river). Soil root-like channels are clearly apparent (vertically oriented to the initial rock layering).

platings are often formed in arid landscapes in the given eluvium. We studied calcareous and gypsic weathering crusts in the composition of ancient intraformational arid eluvium of Tianshan, so-called by the composition of the plating crowing them.

Intraformation Calcareous Weathering Crusts on Sedimentary Cover. These weathering crusts, with a total eluvial profile thickness up to 3-5 m, are widely developed in upper Cretaceous-Eocene and Miocene deposits in a series of Tianshan basins. They are generally formed on a clay, sand-clay, and gruss-clay sedimentary substrate, and they are crowned by dense calcareous plating up to 2-3 m in thickness. The latter are composed of unique lumpy non-layered limestones (Fig. 4, a, b), pelitomorphized or finely crystalline with abundant veins or tubular channels (often in the path of ancient plants), filled by more uncrystallized calcite, with a principally subvertical orientation in relation to the layering surfaces of the rock. Similar carbonate crusts are vividly represented, for example in a section of the Sulutereksk suite of Northern Tianshan (Toruaigyr, Suluterek, Dzhetoyoguz, etc.), and the Irtashsk suite of Western Tianshan (Kyzylsu, Angren, etc.). These crusts have similar aspects and a uniform structure of the eluvial profile.

As an example we will consider the structure of a Miocene profile of calcareous weathering crust in a section of the Irtashsk suite at the Kyzylsu river in the Pretashkent region. Here on the cliffs of the right bank of the Kyzylsu river, the Irtashsk suite is represented principally by red-bed carbonate clays with lenses and interlayers of sandstones, and contains as noted in [5], three uniform plate forming horizons of calcareous crust.

Red-brown non-layered clays with rare calcareous nodules made up of pelitomorphic calcite (the clay facies of proluvium) serve as the maternal rocks of the lower eluvial profile.

In zone I (carbonate-clay), the quantity of calcareous concretions sharply increases; they acquire a principally vertical orientation (in relation to the initial layering of rocks). Tubular segregations of crystallized calcite often occur. They are oriented along the roots of ancient plants, thinning and splitting to the bases of the profile. The thickness of the zone is 1-2 m.

In zone II (the lumpy calcareous plate), the calcite nodules merge into one another and make a dense layer-forming body, made up of pelitomorphic calcite. They have a general greyish-white color (see Fig. 4b). In the rocks of the plating, typical soil textures are distinctly apparent: they are vertical in relation to the layering cleavage, tubular root-like channels (along roots of ancient plants). These channels are generally filled by crystallized calcite or remain partially filled. At the tops of the profile the given channels often are filled by red-brown clay, added from the overlying clay sequence. The thickness is 2.5-3 m.

Above the sharp contact lie red-brown clays of a new cycle of accumulation of the Irtashsk suite, crowned by carbonate crusts similar to those characterized above.

In sections of the Sulutereksk and Irtashsk suite calcareous crusts form both individual horizons and merging packets of carbonate rocks up to 10 m thick. It is noted [5] that locally the thickness of merging calcareous crusts reaches 40-50 m. The most important mineral forming process in the calcareous weathering crusts is the accumulation of calcite in the form of veins and nodules, with formation of dense calcareous plating in the cover of the profile, where metasomatic replacement of initial clay rocks by calcite occurs.

Formation of the given eluvial rocks took place in ancient arid soils and was accompanied by lateral hydromorphic introduction of calcite into the profile from soil-ground water. Neogene soils of the adjacent plain regions of Central Asia are considered in [10]. Two different successively replacing one another formation stages (automorphic and hydromorphic) are often reflected in the thickest varieties of the characterized eluvium (up to 4-5 m). In the automorphic stage, on relatively uplifted parts of dry accumulative plains, formation of deep vertical fractures and coarse deep rootlike channels took place, probably in ancient soils. Subsequently by reduction of relief (under conditions of burial of the accumulative regions) in the subsequent hydromorphic stage of eluvium formation, the lateral input of calcite into the profile from soil-ground waters is enhanced, leading to formation of calcareous plating in initially-porous rocks of the eluvium cover.

Intraformational Gypsic Weathering Crusts. The given crusts from 3-5 m thick are correlated to late Eocene and Miocene deposits, respectively the Sulutereksk and Serafimovsk suites of Northern Tianshan. They are clearly represented, for example in sections of Toruaigyr, Sulutereksk, Serafimovsk, etc.

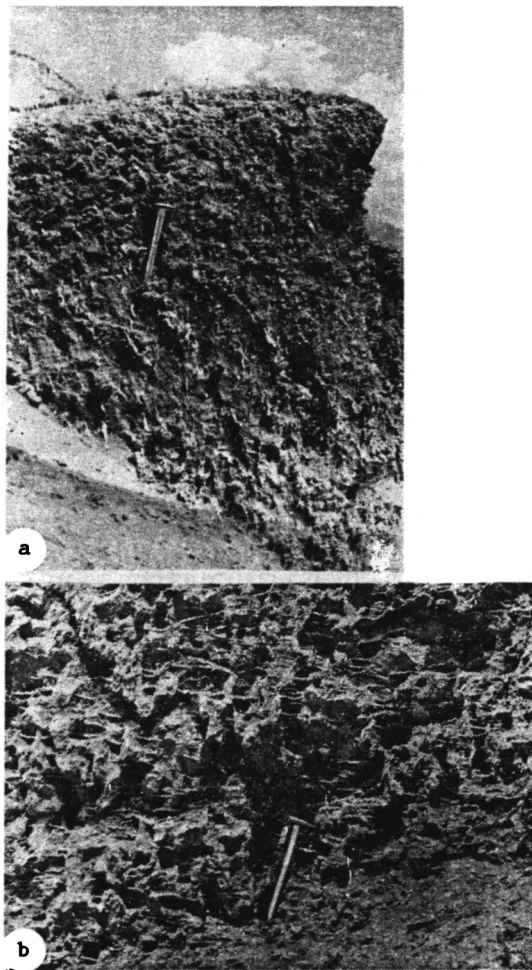


Fig. 5. Intraformational gypsum weathering crust in red-bed clays of the Sulutereksk suite. a) in a section of Toruaigyr of the Issyk-Kul'sk basin gypsum veins with subvertical orientation are seen in the clay in the lower part of the profile; clay-gypsum plating with a fine network of gypsum veinlets is developed in the upper part); b) in a section of Sulutereksk of the Chuisk basin (clay-gypsum plating with wide gypsum veins with subvertical orientation, and with thin horizontal veinlets emphasizing the initial layering).

These crusts are formed on essentially clay or sand-clay surrounding rocks of the above mentioned suites, and they are crowned by clay-gypsum plating (Fig. 5), up to 3-5 m thick.

In the formation of similar eluvium,* a network of intersecting veins, veinlets, and spots of crystalline gypsum is developed in the initial substrate. Gypsum segregations with subvertical orientation in relation to the initial layering predominate among them. These intersect the rarer and thinner veinlets which have a horizontal trend, in agreement with the layering.

The quantity of calcium sulfates (in the clay-gypsum plating) increases towards the tops of the eluvial profile and reaches 50-70% of the volume of the initial rocks (see Fig. 5). Mutual intersection of the vertical and horizontal veins of crystalline gypsum (from 0.5 mm to 5 cm thick) causes formation of a characteristic cellular texture of the rocks in section. In plan, the gypsum segregations most frequently fill a polygonal system of fractures, reminiscent of takyr. In addition, tubular segregations of crystalline gypsum are observed (Fig. 6). They fill cavities from roots of ancient plants.

*The term "eluvium" is employed in the sense used by geologists, not by soil scientists.

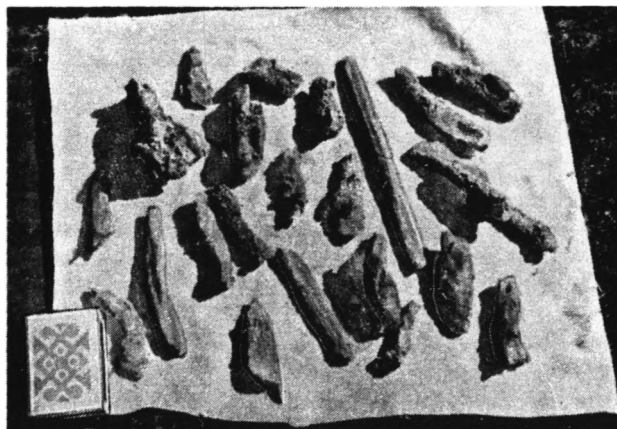


Fig. 6. Tubular segregations of gypsum in voids from ancient plant roots (from clay-gypsum plating of intraformational weathering crusts in the Sulutereksk suite of a section of Toruaigyr).

Abundant root-like channels or Takyr fractures, filled by gypsum, make it possible to consider that formation of the given eluvium took place through active participation of ancient arid soils (takyr and salt-bottom). Formation of takyr fractures and deep root-like channels probably took place in the automorphic stage of eluvium development, and subsequently salt-bottom soils were subsequently formed (with the decrease in relief) in the hydromorphic stage. This was accompanied by the lateral input of gypsum into the profile, and the filling of fractures, voids, and led to the generation of clay-gypsum plating in the most porous, water-permeable upper parts of the eluvium.

In the characterized sections, the variously aged gypsum crusts (similar to the calcareous crusts characterized earlier) may merge into a single clay-gypsum plating up to 10-12 m thick. In the structure of similar plating it is often possible to see compound gypsum crusts, distinguished from one another by color, gypsum saturation, and size of the gypsum cells formed by interaction of the gypsum veins (from 2 to 10 cm).

Textural transformation of the initial substrate, its gypsification with generation of gypsum plating in the upper part, and metasomatic replacement of clay by gypsum are the main mineral-forming processes in the considered eluvium.

In Cretaceous-Paleogene and Neogene deposits of adjacent regions of Kazakhstan and Central Asia, the formation of gypsum crusts similar to Tianshan with the participation of takyr and salt-bottom soils is considered in a series of publications [11, 14, 18].

In conclusion we note that in upper Cretaceous, Paleogene, and Neogene sections of Tianshan, platings with dolomite, gypsum-siliceous, or gypsum-carbonate-siliceous mineralizations occur in the composition of intraformational arid eluvium. Their study is a problem for further investigations. The question of possible authigenic clay formation in various types of intraformational eluvium, including the varieties characterized above, also requires special consideration. Beyond the limits of Tianshan, judging by a series of publications [2, 11, 14, 18, 29, 30], formation of palygorskite, smectites (aluminum and super-dispersed), inter-layered kaolinite-montmorillonite formations, etc., was recorded in Mesozoic and Cenozoic arid intraformational eluvium.

OVERALL REGULARITIES IN THE LOCALIZATIONS OF ANCIENT ARID ELUVIUM IN PALEO-LANDSCAPES

Analyzing the results of lithologic-facies and paleo-floristic investigations in the characterized regions of Tianshan [1, 11], it is possible to propose that in the late Cretaceous-Eocene and Miocene stages their development here governed arid landscapes: dry savannahs, deserts, and semi-deserts. Denudational landscapes were leveled or very weakly separated, and accumulative landscapes represented alluvial-proluvial or salt-bottom plains with separate lakes, often saline.

Utilizing regularities in the structures of recent eluvial cover in arid regions, we note a general regularity in the distribution of separate types of arid weathering crusts. Gruss and gruss-clay types of eluvium (on crystalline rocks of basement) which are generally weakly carbonate or gypsum dominated in ancient water divides in autonomic landscapes. Lower, in the zone of transition to an accumulative region (in the super-aquatic landscape type), calcareous or gypsum plating often formed in the upper part of the characterized weathering crust. Still lower, in accumulative landscapes on a sand-clay substrate, intraformational eluvium formed, crowned by calcareous or gypsum plating. The lowest position is occupied by weathering crusts with gypsum plating, which gravitate to strongly hydromorphic lacustrine-salt-bottom parts of paleorelief.

1. Miocene and particularly clearly expressed Maastrichtian-Eocene epochs of strengthening in the intensity of arid chemical weathering and formation of different types of infiltrated eluvium are recorded in the late Cretaceous and Paleogene-Neogene history of geologic development of Tianshan.

2. Arid weathering of the substrate often took place with the active participation of ancient soils, and it was accompanied not only by physical, but also chemical transformation (argillizations of aluminosilicates, appearance of calcareous or gypsic metasomatism, etc.).

LITERATURE CITED

1. Atlas of Lithologic-Paleogeographic Maps of the USSR [in Russian], Vol. 3, 4, GUGK (1967).
2. G. I. Blom, "Buried palygorskite soils in lower Triassic deposits of the Moscow syncline," Dokl. Akad. Nauk SSSR, 194, No. 2, 295-398 (1970).
3. B. A. Bogatyrev, "Bauxites of activated parts of the Scythian and Turansk plates," in: Bauxite-Bearing in the Main Tectonic Structures of the Earth's Crust [in Russian], Nauka, Moscow (1988), pp. 147-153.
4. L. N. Bertunov, Lithology, Paleogeography, and Problems of Oil and Gas Bearing in Cenozoic Molasse Formations of the Issyk-Kulsk Basin, Dissertation in Geol.-Min. Sci., Tashkent State Univ., Tashkent (1969).
5. V. P. Volkov and B. I. Pikhasov, The Geology and Hydrogeology of the Upper Cenozoic Holodnoi Steppe and Eastern Kyzylkums [in Russian], Fan, Tashkent (1985).
6. R. Garrels and F. MacKenzie, Evolution of Sedimentary Rocks [Russian translation], Mir, Moscow (1974).
7. I. I. Ginzburg, "Types of ancient weathering crusts, their forms, appearance and classification," in: Weathering Crusts [in Russian], No. 6, Izd. AN SSSR (1963), pp. 71-101.
8. B. P. Gradusov, "Genesis of palygorskite in continental and oceanic sediments," Dokl. AN SSSR, 230, No. 2, 418-421 (1976).
9. B. P. Gradusov, "Clay minerals in soils of desert regions," Thesis Lectures of VI Delegation of Convention of the All-Union Association of Soil Scientists, Tbilisi, 133-134 (1981).
10. V. V. Dobrovolskii, "Hypergene formations of Eastern Africa," in: The East-African Rift System [in Russian], Vol. 2, Nauka, Moscow (1974), pp. 5-56.
11. V. S. Erofeev and Yu. G. Tsekhovskii, Paragenetic Associations of Continental Deposits. Family of Arid Paragenesis. Evolutionary Periodicity [in Russian], Nauka, Moscow (1983).
12. I. D. Zkhus, Clay Minerals and Their Paleogeographic Value [in Russian], Nauka, Moscow (1986), pp. 5-278.
13. N. S. Kasimov, The Geochemistry of Steppe and Desert Landscapes [in Russian], Izd. MGU, Moscow (1988).
14. N. S. Kasimov, Yu. G. Tsekhovskii, B. A. Gradusov, et al., "Miocene fossil soils of Kazakhstan," Litol. Polezn. Iskop., No. 2, 3-19 (1983).
15. M. R. Lider, Sedimentology [Russian translation], Mir, Moscow (1986).
16. B. M. Mikhailov, "Hypergenesis in the arid tropics of East Africa," Litol. Polezn. Iskop., No. 5, 3-26 (1970).
17. B. M. Mikhailov and G. V. Kulikova, Facies Analysis of Weathering Crusts [in Russian], Nedra, Leningrad (1977).
18. A. I. Perel'man, Molar Migration Processes on the Plains of Eastern Turkmeniya and Western Uzbekistan in the Neogene [in Russian], Izd. AN SSSR, Izd. AN SSSR, Moscow (1959).
19. B. A. Petrushevskii, "Structure of Tertiary Deposits of Tianshan," Byul. MOIP. Otdel. Geol., 13, No. 1, 53-91 (1948).
20. Soils in Ancient Weathering Crusts [in Russian], VASKhNIL, Moscow (1979).
21. M. A. Rateev, "On the relations of allothigenic and authigenic clay formations in different types of lithogenesis," Litol. Polezn. Iskop., No. 2, 39-62 (1964).

22. M. B. Rukhin, "Weathering, transport, and deposition of material," in: Guiding Handbook of the Petrography of Sedimentary Rocks [in Russian], Vol. 1, Gosnauchtekhizdat, Leningrad (1958), pp. 40-61.
23. V. M. Sinitsyn, Introduction to Paleoclimatology [in Russian], Nedra, Leningrad (1980).
24. A. V. Sochava, "Carbonate crusts (caliche) in red-bed formations of the Precambrian and Phanerozoic," in: Facies and Geochemistry of Carbonate Deposits [in Russian], VSEGEI, Leningrad (1973), pp. 72-73.
25. A. V. Sochava, "Red-Bed formations of the Precambrian and Phanerozoic [in Russian], Nauka, Leningrad (1967).
26. N. M. Strakhov, Basic Theory of Lithogenesis [in Russian], Izd. AN SSSR, Moscow (1960).
27. V. N. Troitskii, Upper Triassic and Jurassic Deposits of Southern Uzbekistan [in Russian], Nedra, Leningrad (1967).
28. A. T. Turdukulov, Paleogene and Neogene Geology of Northern Kirgizia [in Russian], Ilim, Frunze (1987).
29. Yu. G. Tsekhovskii, B. P. Gradusov, and N. P. Chizhikova, "Process of mineral formation in siliceous-calcareous profiles of soil forming and weathering from the paleo-savannah of Eastern Kazakhstan," *Litol. Polezn. Iskop.*, No. 2, 50-66 (1973).
30. Yu. G. Tsekhovskii, B. P. Gradusov, and B. P. Chizhikova, "Mineralogic composition and structural features in buried soils of the Zimunaisk suite of Eastern Kazakhstan," *Dokl. AN SSSR*, 219, No. 6, 1461-1464 (1974).
31. A. M. Tsekhomskii, "Weathering crusts," in: Lithology Handbook [in Russian], Nedra, Moscow (1983), pp. 244-251.
32. L. G. Chernyakhovskii, "Climatic zonality of the eluvial process," in: Processes of Continental Lithogenesis [in Russian], Nauka, Moscow (1980), pp. 28-60.
33. I. S. Chumakov, Cenozoic of the Metalliferous Altai [in Russian], Nauka, Moscow (1965).
34. S. S. Shul'ts, Analysis of the Most Recent Tectonics and the Relief of Tianshan [in Russian], Geografiz, Moscow (1948).
35. V. N. Shcherbina, Mineral-Petrographic Features of Tertiary Continental Salt-Bearing and Gypsum-Bearing Deposits of Intermountain Basins of Tianshan [in Russian], Izd. AN KirgSSR, Frunze (1956).
36. V. M. Yazovskii, "Stratigraphic scheme of Paleogene-Neogene deposits of Northern Kirgizia," in: Northern Tianshan in the Cenozoic [in Russian], Ilim, Frunze (1979), pp. 3-16.

LANDSLIDES AND THE FORMATION OF HOMOGENEOUS HORIZONS IN SUBMERGED DEBRIS CONES (LIASSIC, NORTHEASTERN CAUCASUS)

Yu. O. Gavrilov

UDC 551.435.627:551.762(479)

Sediments formed in an area of ancient debris cones are characterized by a high sedimentation rate, showing that gravitational processes, marked by various types of landslide, were important in their development. In one type of gravitational process, major slices of sediment moved downslope along a number of discrete disruption planes within the section. The originally bedded sediments in these intervals were typically displaced and worn down, leading to the formation of homogeneous rock horizons. During lithification of these horizons, lateral stresses caused fracturing in certain rocks. This is regarded as lithification cleavage.

Investigation of modern sedimentation in regions of fluvial debris cones shows that the high rates of sediment accumulation on the sloping underwater sediment surface lead to unstable sediment masses and the onset of landslide processes. Moreover the scale of these phenomena may vary from quite minor features to major slide blocks tens of kilometers long [1, 2, 4, 7, and others].

Geological Institute, Academy of Sciences of the USSR (GIAN SSSR), Moscow. Translated from *Litologiya i Poleznye Iskopaemye*, No. 1, pp. 49-58, January-February, 1990. Original article submitted May 16, 1989.

During a study of sediment accumulation in the Early and Middle Jurassic basin of the Northeastern Caucasus, in the zone of development of sediments in the submerged debris cone of an ancient delta (Dagestan), we repeatedly noted horizons which were formed in connection with various types of sediment landslide. Two types of landslide process may be distinguished here. One of these comprises sediments generated from the disruption of accumulations of sands and argillaceous sands on sloping surfaces. Their rapid movement and simultaneous mixing led to the formation of horizons with chaotic structure. In this type of landslide the underlying sediments are partially eroded, but the earlier rocks are virtually unaffected by deformation.

The other type of landslide consists of the downslope movement of enormous and very long plates of sediment, occurring without appreciable disruption of their internal structure. At the same time these landslides, being generally of a relatively large scale, are accompanied by the deformation of some of the sedimentary intervals over which the sedimentary plates have moved. In this paper we investigate the structure and likely mechanism of formation of these deformed horizons.

Study of the distribution of this type of horizon in the upper Lias shows that their position in the succession follows a clear pattern. They are absent in dominantly clayey intervals from tens to hundreds of meters thick, but they occur within deposits forming complexes with an alternating rhythmic structure. Figure 1 depicts one of these rhythms, typical of the Jurassic succession in this area. The rhythm was generated by deposition from one of the prograding sand lobes of the submerged debris cone. Sections of terrigenous rock bodies with similar structure, and accumulated in analogous situations, are described in a series of studies [8, 10 and others].

In rhythmic succession, horizons of deformed sediments are encountered within intervals of interbedded argillaceous and sandy beds and lenses. The thickness of these horizons varies from several decimeters to several meters, occasionally reaching 10 m and more.

On what does the occurrence of the deformed horizons precisely within intervals of interbedded sandy claystone depend? To answer this question it is necessary to study some particular conditions of sedimentation.

Evaluation of the speed of formation of different parts of the rhythms shows that the rate of accumulation of the rhythmic deposits did not remain constant. Minimal rates were characteristic of the lower parts, which are predominantly composed of the greatest thicknesses of claystone-siltstone sediments. At the time of formation of higher levels of the rhythms, the rate of sediment accumulation had generally increased on account of the appearance of sand beds, formed as a result of the introduction of sandy material in quite shortlived flows. Finally, the maximum rate of sediment accumulation was characteristic of the topmost section of the rhythms with a fair thickness of sandy layers. Such a progressive increase in the rate of sediment accumulation led in some cases to the natural processes of sediment compaction (particularly of clays), with the accompanying expulsion of pore fluids, not proceeding to completion. The zone of sediment lithification was therefore extended and the sediments, being water-bearing, remained for a long time in a plastic condition. The presence of sandy beds in the succession also appreciably complicated the process of expulsion of water from the clays. It is evident from the pattern of water migration from the clays in the succession (Fig. 2) [6], that the water was primarily expelled in sand beds, from the immediately adjacent sections of clay beds. At the same time, the central parts retained a considerable water content, and therefore remained in a more plastic condition.

The deposition of thick sandstone horizons at the top of the rhythms led to a significant increase in the overburden, and the pressure on underlying rocks. Since sedimentation in the study area occurred on the sloping surface of a debris cone, gravitational forces began to operate in a tangential direction. These caused displacement along surfaces in the center of some rhythms, and the onset of movement of sedimentary sheets, composed of the upper parts of the rhythms, down the slope. This process may in some cases have been initiated by earthquakes, but generally all the requirements were present for it to develop independently.

The actual structure of the deformed horizons, together with their texture and thickness, vary within wide limits. They depend on such factors as the nature of the original sediments, the degree of lithification at the time when deformation began, and the extent of the phenomenon, which in turn determines the thickness of the slid sheet, and the distance over which it moved.

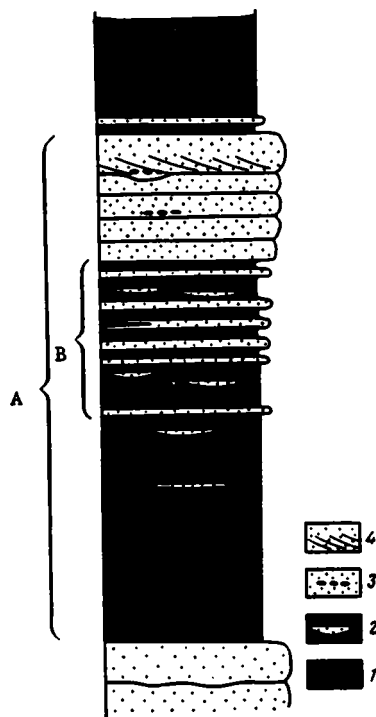


Fig. 1

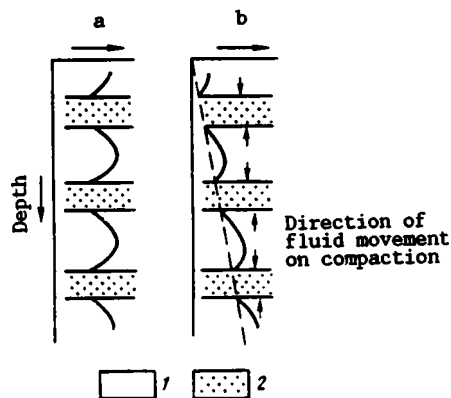


Fig. 2

Fig. 1. Section through part of a sandy-argillaceous unit within the debris-cone deposits. [A) Structure of one of the rhythmic units; B) most dense interval of disruption surfaces.] 1) Clayey-silty deposits; 2) sand-filled erosional scour formed by temporary bottom-current; 3) sandstones with intercalations of lenticular concretions, washed out of clays; 4) sandstones with unidirectional cross bedding.

Fig. 2. Pressure distribution in buried water in the interbedded sandstone-claystone rock unit [6]. a) Claystone porosity; b) distribution of fluids in claystones. 1) Claystones; 2) sandstones; dotted line denotes hydrostatic pressure.

A diagnostic sign which is valid for practically all the horizons where mixing has developed is the partial or complete disappearance of original sedimentary bedding. Their distinctive appearance is apparent when they are compared with the under- and overlying rocks, in which interbeds of sandstone and clayey siltstone are developed. The disappearance of the interbeds is related to the plastic condition of the deposits within the horizons in which movement occurred. The movement gradually wore down the boundaries between the beds, which began to be indistinct, and finally to disappear completely. As a result of this wearing down of the beds, the original grain-size variation was lost. Diagenetic concretions, which are often encountered in adjacent beds, suffered an analogous process. At the time of onset of deformation they had not been consolidated and, being in a plastic state, were extended and dispersed throughout the rock.

In thin section it is evident that in rocks from deformed horizons, fragments of components of various grain sizes are present, so that they have the appearance of poorly sorted sediments. Mica flakes and other grains are arranged in a disordered manner, whereas in adjacent beds in the succession, unaffected by movement, platy terrigenous particles are usually arranged parallel to bedding.

It should be noted that on exposed surfaces of these horizons, interbedded bands, usually lenticular, occur. They are distinguished by slight changes in color, or the nature of the weathering. This banding appears to be a relict indication of former interbedding.

So the main result on the lithology of this process occurring in the studied horizons is the homogenization of the sediments. In different places it is present to a greater or lesser degree, but in essence it is always the same. We propose the term homogeneous horizons for these formations, which we will use in the following account.

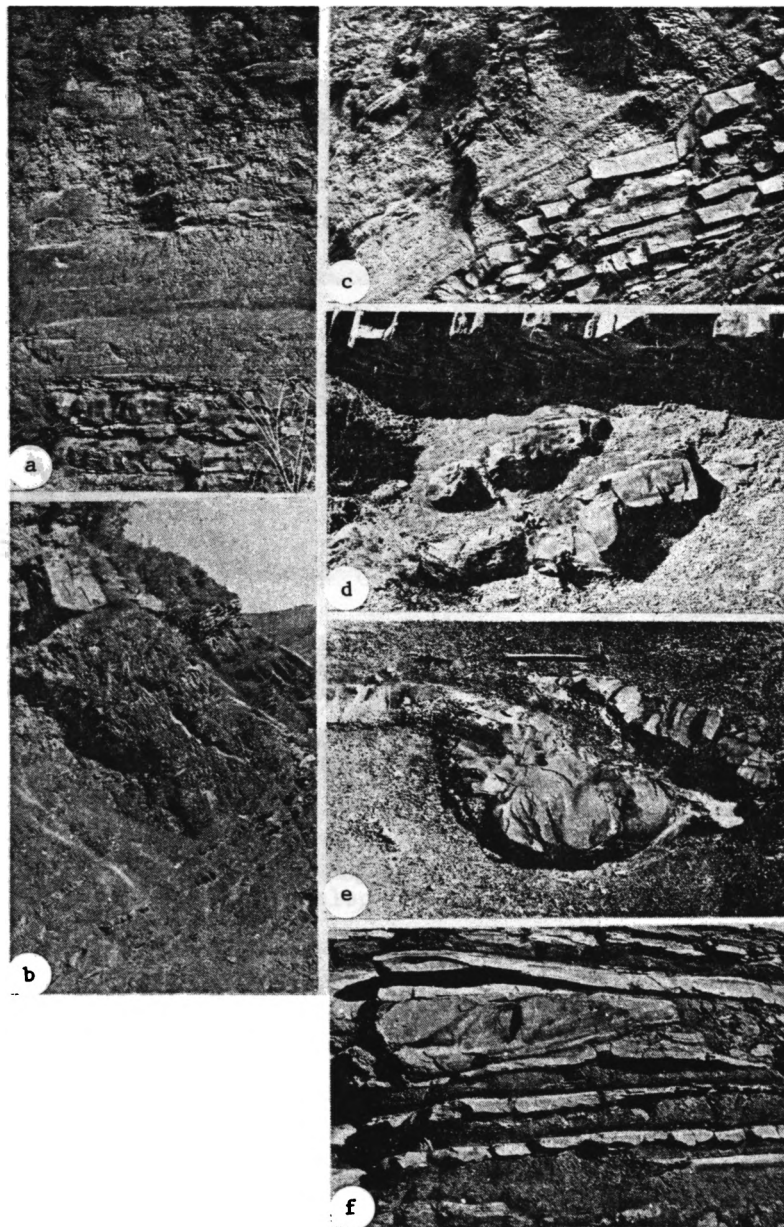


Fig. 3

As has been noted, the degree of homogenization of the sediments depends on a combination of factors. Related to this are horizons which vary markedly in the degree of reworking of primary structures, as indicated by their present structure.

In the most homogenized horizons, no original structures remain, and only on some intervals is weak banding observed (Fig. 3, a, b).

Sandstone interbeds in some units were already quite well consolidated when landsliding began, owing for example to early diagenetic carbonate cementation. In this case they did not undergo the process of wearing down and mixing with clayey-siltstone material, but were subject to boudinage. On Fig. 3c, for example, a series of thin lenses are apparent lying in the lower part of the homogeneous horizon, formed from an earlier sand bed. Sometimes, in echelon fragments of sandstone beds are encountered, formed by the stacking of one lens on another (Fig. 3, a, b). In some places, the deformation and wearing down of quite major sand bodies is observed. In Fig. 3f, a sedimentary sandstone lens can clearly be recognized in a homogeneous horizon. It is faulted in the center and to the right it thins.

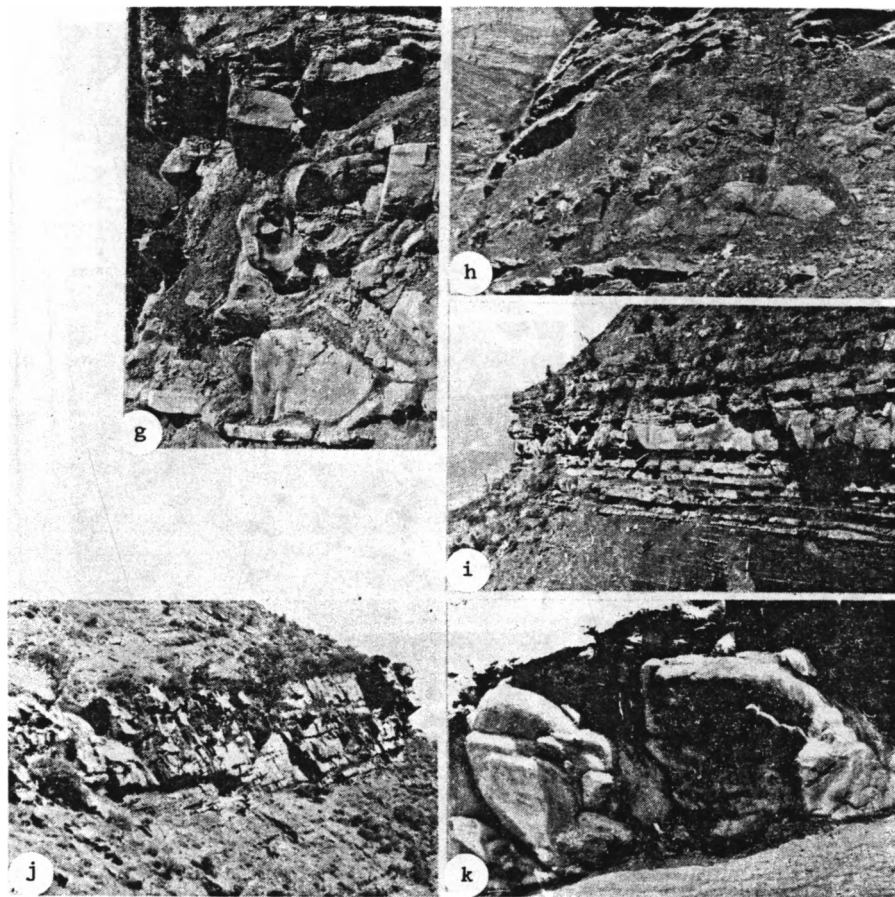


Fig. 3. Homogeneous horizons (HH) with various structural features and rock compositions, in various stages of deformation. a-b) HH with weak banding caused by relict grain-size inhomogeneities in the rocks; underneath lie unaffected rocks with original sedimentary bedding; c) HH with sandstone lenses, the remains of sedimentary sandstone layers; d-e) an echelon lenses of sandstone within a homogeneous clayey siltstone matrix; above is the base of the overlying sandstone bed; f) deformed and smeared sedimentary sandstone lenses lying within a HH; g) sandstone body with complex structure, from HH with various types of deformation; above is the base of a thick sandstone bed; h) HH, the lower part of which is characterized by deformed sedimentary sandstone beds, contained within a mass of sandy siltstone of similar composition, and above with a more clay-rich composition; i) HH with deformed sandstone lenses below a massive sandstone bed; j) truncation of the lower beds of a sandstone horizon by its movement down the sloping surface of a debris cone; k) evidence of basal intrusion of pressured sandstone material into the underlying clayey-siltstone mass.

Deformation of some sandstone beds led occasionally to the development within the homogeneous horizon of considerable disorder, and structurally complex sandstones, in which both plastic and brittle deformation are observed (Fig. 3g).

In those cases where different lithologies were originally present in homogenized horizons, later mixing of material may not have affected the entire thickness of these horizons, but its composition would differ from that in the underlying horizon. Figure 3h shows a homogeneous horizon subdivided into a lower, sandy, and an upper, relatively clay-rich section. In the upper part complete mixing of the sediments occurred, but in the lower, lenticular fragments formed from the former sandstone beds, which were then cemented within a sandy siltstone, which did not differ significantly in composition from the included lenses. These lenses therefore have no distinct boundaries, and a smooth transition is often observed between them and their host rocks.

Disrupted horizons are repeatedly observed in direct contact with the bases of massive sandstone beds forming the top of a rhythmic unit (Fig. 3i). Boudinage is also observed in the

underlying bed, and sometimes convolute inclusions of sandstone. A feature of these intervals for determining the type of deformation may be found at the base of the sandstone horizons, where there are furrows thought to be extensional fissures. Sandstone horizons are generally composed of a series of beds. As a result of the expulsion of water from clays, a lower bed or series of beds became the most highly water-saturated, which in a number of cases promoted its thinning and shearing during the process of downslope transport [Fig. 3j]. Such a relationship with the underlying deposits could mistakenly be taken for erosional truncation by a short-lived underwater current. Note also that the thinned sandstone mass at the basal part of the sand beds was sometimes intruded under pressure into the underlying clayey-siltstone mass, giving rise to characteristic signs of intrusion at the base (Fig. 3k), directed downslope.

Formation of Lithification Cleavage. One of the characteristic signs of the homogeneous horizons is the development within them of subvertical fracturing (Fig. 4). It is most intense in the thickest horizons. It is particularly noticeable that the fractures do not continue into the adjacent rock units, which retain their original bedding. Fracture surfaces are mostly parallel to one another, although they are seldom planar. They are generally curve, and sometimes wavy. As has been noted, faint relict banding can be recognized in homogeneous horizons, caused by slight grain-size variations. This factor clearly explains the variable direction of the fracture planes. An analogous feature is observed in sandy shale beds in which tectonic cleavage is developed, the planes of which change their orientation in passing from argillaceous into sandy beds.

The separation of individual fracture planes varies from a few millimeters to 1-1.5 cm, and depends on the lithological composition of the rock. A tendency for a reduction in the separation in more argillaceous rocks is noted, where the proportion of sand-silt material is slight.

An important factor in understanding the cause of the fracturing is its geometry. Measurements of fracture orientation show that they range in direction from 130-150° N, towards the southeast. This is virtually perpendicular to the direction of the short-lived currents down the slope, and the direction of landsliding of sedimentary beds. They are therefore virtually parallel to the strike of the slope of the sedimentary surface in this part of the Liassic basin.

Taking these factors into account, the formation of the fracturing seems to have occurred in the following manner. The initiating of the landsliding process was largely caused by the increasing water content of some interbedded sandstone-claystone intervals, which was assisted, as noted above, by the well-stratified nature of the deposits which hindered the expulsion upward of water from clays. The landsliding of sedimentary sheets, and the formation of homogeneous horizons, where stratification was lost, removed this obstacle, and led to the expulsion of formation waters, and dehydration of the lithologies. As a result, the viscosity of the sediment in the horizons on which the sliding body moved increased, leading to the cessation of movement. However, although the material in the homogeneous horizons had lost its plasticity, it continued to be under the influence in a tangential direction of the stresses down the slope. Moreover, with the further deposition over a period of time of sediments on top of the transported plate, these stresses increased. However, the increased viscosity of the sediments prevented further movement of the slid block.

Therefore, the lithification of the homogeneous deposits occurred under the influence of lateral stresses, which led to a stage of postsedimentary transformation, with the generation of a system of unidirectional fractures, orientated perpendicular to the prevailing stress direction. Fracturing of this nature is essentially similar to tectonic cleavage. However, we consider that it has all the features characteristic of late diagenetic lithification cleavage.

As noted by Lomtadze [3], the origin of cleavage may be endogenic, arising from processes deep in the earth's crust, or exogenic, encountered quite rarely and related to surface effects. The cleavage examined here may clearly be assigned to the exogenic category, although it formed not in the highest beds in the succession, but at some depth.

The movement downslope of the slid blocks and the formation of homogeneous horizons was apparently not a unique and short-term process. Its origin should be related to the deposition of the sandstone beds at the top of the sedimentary rhythm, and the minimum depth at which shearing occurred along some surfaces may have been about 4-6 m. Homogeneous horizons

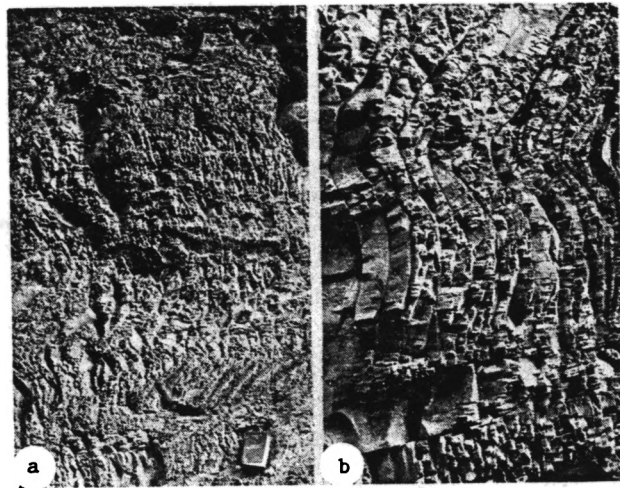


Fig. 4. Lithification cleavage. a) General view of part of a HH with intense development of cleavage (size of diary is 10×15 cm); b) detailed structure of cleavage in rock.

generally formed at depths of 10-30 m (lithification cleavage forming, apparently, at some considerable depth). At this stage of the postsedimentary transformation, argillaceous beds on which shearing initially occurred were characterized by a certain combination of mechanical properties. In accordance with the engineering-geological classification of argillaceous rocks, they corresponded to a great extent with rocks of group II [3], to which are usually referred soft, poorly consolidated varieties with viscous, cryptoviscous or plastic consistencies. The molecular interaction between the clay minerals has been weakened by the development of films of aqueous or aqueous-colloidal salt-bearing films, which have a plastifying (softening) effect. Under slight pressure, rocks of this group exhibit some shear strength, displaying elastic properties, but under increasing pressure they rapidly develop plastic deformation.

It should be emphasized that in the type of landslide under investigation, movement down the slope of the debris cone does not occur as individual small blocks, but as quite extensive sheets. Landslides occur over a broad front, possibly extending from many hundreds of meters to kilometers, and larger dimensions should not be ruled out.

The distance over which sediment sheets were transported is difficult to evaluate. It should just be noted that homogeneous horizons may be generated even by processes operating on a relatively small scale.

Variations in the thickness of individual homogeneous horizons may be noted. This may presumably result from the compression and redistribution of the pressurized plastic mass within the horizon, or simply to the doubling of thickness due to the thrusting of one part of the horizon over another.

The processes under examination contributed to the development of landslides in other rocks. Sometimes, fragments of the transported sheet (of various sizes) became detached from its frontal zone, and moved down the slope in a quite different manner, with rapid displacement and mixing of the sedimentary mass. As a result, sediments were generated composed of fragments of argillaceous and sandy rocks, cemented with sandy material. This type of sedimentary formation will be examined in a future paper.

In areas of the slopes of modern oceans and marine margins, various landslide formations are recorded, which reach quite considerable dimensions [2, 4, 7, and others]. In particular, the displacement has been established of large-scale sedimentary sheets, the extent of which may measure many kilometers, and even tens of kilometers (with certain slides of hundreds of kilometers), whereas their thickness is comparatively slight (a few tens to hundreds of meters). The landslides from the Liassic of the Northeast Caucasus appear broadly similar to the modern examples. It should therefore be expected that the beds on which movement occurred in recent times would be similar to the homogeneous horizons recorded here. The direct observation of beds underlying modern landslides is difficult. However, there is some indirect evidence for their presence, particularly from seismic profiles. Various researchers [9 and others] have noted that the interior of major slid blocks on seismic profiles often

resembles parts of the surrounding continental slope, but below them are layers with far poorer acoustic characteristics, containing almost no reflecting surfaces. These beds appear to be intensely deformed homogeneous horizons. Moreover, taking into account the diverse morphologies noted in the homogeneous horizons from ancient sediments, the various beds on which modern slid blocks have moved may appear on seismic profiles in a variety of different forms.

Studies of modern oceans also show that landslides arise in very diverse conditions, including slopes with a very gentle gradient [1, 5], and with relatively low sedimentation rates [9]. However, in these situations the landslides may be rare chance events, triggered by external causes such as earthquakes, hurricanes, etc. The appearance in the sedimentary succession of analogous homogeneous horizons therefore also occurs sporadically.

Particular features of the Liassic homogeneous horizons examined are their confinement to a specific region within the paleobasin, and their frequent appearance of a set position within the sedimentary sequence. This suggests that they were typically formed in specific sedimentary environments. In such environments, the conditions necessary for landslides to occur did not arise by chance, but as a natural extension of the sedimentary process. In fact, examination of coeval rock bodies in other regions of the Northern Caucasus shows that homogeneous horizons elsewhere are very rare, and usually absent. It therefore seems to us that the appearance in a section of a series of homogeneous horizons may be considered when reconstructing sedimentary environments in a paleobasin to be diagnostic of an ancient debris cone.

LITERATURE CITED

1. A. D. Arkhangel'skii, "Landsliding of sediments on the floor of the Black Sea and its geological significance," Byull. Mosk. Ova. Ispyt. Priro. Otd. Geol., 8, No. 12, (1930).
2. A. P. Lisitsyn, Sedimentary Avalanches and Sedimentary Breaks in Seas and Oceans [in Russian], Nauka, Moscow (1988).
3. V. D. Lomtadze, Engineering Geology. Engineering Petrology [in Russian], Nedra, Leningrad (1984).
4. V. N. Moskalenko and K. M. Shimkus, "The role of landslides in the formation of olistostromes and olistostomes in the Lower Cenozoic deposits of the Black Sea," Okeanologiya, 16, No. 4, 655-661 (1976).
5. K. B. Lewis, "Slumping on a continental slope inclined at 1-4°," Sedimentology, 16, No. 1/2, 97-110 (1971).
6. K. Magara, Compaction and Fluid Migration, Elsevier, Amsterdam (1978).
7. S. Saxov and J. K. Nieuwenhuis (eds.), Marine Slides and Other Mass Movements, Plenum Press, New York (1982).
8. E. Mutti and F. Ricci Lucchi, "Turbidites of the Northern Apennines: introduction to facies analysis," Internat. Geol. Rev., 20, No. 2, 125-166 (1978).
9. C. P. Summerhayes, B. D. Bornhold, and R. W. Embley, "Surficial slides and slumps on the continental slope and rise of South West Africa: a reconnaissance study," Marine Geology, 31, No. 3/4, 265-277 (1979).
10. R. G. Walker, "Deep-water sandstone facies and ancient submarine fans: models for exploration for stratigraphic traps," Bull. Am. Assoc. Pet. Geol., 62, No. 6, 932-966 (1978).

Anomalous phenomena examined in bituminous clays (increased temperature, formation and pore pressure, inhomogeneities in the density and catagenesis of the sediments, etc.), are caused by cavitation processes during the fluid-induced fracturing of the rocks. The formation of reservoirs for oil accumulation and the clay anomalies are linked in time, space and origin. Varying rock density within an argillaceous sequence may provide an indication of the presence of oil.

Problems of establishing the conditions under which fracturing occurs in argillaceous reservoir rocks, and of determining the origin and distribution of oil within the fractures, have deservedly attracted much attention from researchers [2-5, 10, 12, 14, 17, 18, 20, 22-24, 26, 29-32]. Some aspects of these problems, connected in part to interpreting the nature of anomalies in bituminous clays, have considerable scientific and applied significance. Studies have been undertaken on Lower Maikopian clays of the Central Caucasus, supplemented where necessary with data from the literature on the Bazhenovsk Formation of Western Siberia.

In the Maikopian sediments, oil is produced at an economically viable rate from the lower parts of the Batalpashinsk Formation and the upper parts of the Khadumsk Formation (P_3^{1-2}), and extensive oil shows occur in the Beloglinsk and Kumsk Formations (P_2^3) throughout the Eastern Stavropol' Trough at depths of 2000-2200 m.

The Oligocene clay is generally 40-50 m thick, and has horizontal bedding characterized by intercalations of dark, almost black, highly bituminous beds, dark-grey non-bituminous beds, and thin beds of microlaminated rock, from 1-3 mm thick. The variations in color and bedding type depend to a large extent on the precipitation on bedding surfaces of fossil organic matter (OM) of mesostructured character, and also on variations in sedimentation rate. Beds of varicolored clay are structurally and texturally identical and have the same mineralogical composition, indicating that there were no significant changes in the hydrodynamic conditions during the sedimentation of the fine flakey clay mineral aggregates, which have a constant montmorillonite-chlorite-kaolin-hydromuscovite (46-62%) and kaolin (26-40%).

Interbedded with the clays are thin (0.1-5 mm) lenticular laminae and lenses of light-grey poorly sorted siltstones, and also beds of various thickness of limestone, marl, and dolomite.

In the black beds of oil-saturated clays, the average content of residual dark-brown OM is 8% (TOC from 2-9%), so it can be considered as a rock-forming component. Three basic forms of OM are differentiated: dispersed (dominant); threadlike (lenses and veinlets of amorphous matter); and carbonaceous (appearing as silt-sized grains). The general content of bitumen (generally of a parautochthonous type) varies from 0.08-1.24%. Elemental and molecular composition corresponds to a sapropelic kerogen (dominated by alginite microcomponents), which is evidence of the oil source potential of the hydrocarbon (C from 84-88%, H from 10-14%, other elements 1-2%). The catagenesis of OM to the level of oil generation in the sediments is reached, where it is only slightly reduced, in the catagenic grades PC_1^1 to MC_1^1 , but for the dominant reduced OM type, catagenic grades from PC_3^1 to MC_3^1 are required.

The content of organic carbon (C_{org}) is closely correlated with the radioactivity of the rocks (up to 45 $\mu R/h$), owing to the adsorption by autochthonous organic matter of dispersed uranium. In beds of productive clay there are numerous phosphatized traces of fish scales and fragments of fish skeletons with calcite cement, remains of benthonic algae and allochthonous detritus. The organic matter (including that wholly or partly reduced to sulfides) contains authigenic marcasite and pyrite, which comprises 15-18% of rock volume in the oil-saturated zones, which is more than that in the non-bituminous clays.

North Caucasian Natural Gas Scientific Research Institute, Stavropol'. Translated from *Litologiya i Poleznye Iskopaemye*, No. 1, pp. 59-68, 1990. Original article submitted April 4, 1988.

The main phase of oil generation corresponds to the transition of OM from the protocatagenic to the mesocatagenic stage. The destruction of part of the OM leads to the generation of liquid and gaseous hydrocarbons. Hydrodynamic separation of the hydrocarbons is guaranteed by the presence of the OM concentrated in argillaceous units within thick bodies of poorly consolidated and practically impermeable rocks [5]. The oil evolved is of a methanonaphthalene type, not catagenetically altered, with low gas content, and relatively light (density 0.83-0.86 g/cm³). Paraffin content is 4.9-11.9%, sulfur content is low (less than 0.19%), and tar content is 2.3-16.6%. The asphaltene content is characteristically low (0.24-1.27%), and there is a complete absence of water. The lithological and geochemical characteristics of these deposits and formation fluids are quite fully described in [2, 5, 7, 8, 12, 20, 32].

The oil-bearing rocks in the Lower Maikopian are characterized by a series of anomalous phenomena (a relative increase in the background temperature, and in the formation and pore pressures, inhomogeneous rock densities, etc.), which distinguish them from the over- and underlying deposits (Table 1). The nature of these phenomena may be understood by discovering the fracturing mechanism in the clays.

Claystone reservoirs have been well-known for a long time. In the opinion of most researchers, the filtration capacity of such reservoirs ensures that generally horizontal fracturing of the rocks occurs on the meso- and macroscale [2, 4, 5, 10, 12, 14, 18, 20, 22, 24, 26, 27, 29, 30]. Fractures in the Lower Maikopian argillaceous reservoirs are subhorizontal and branching, and of varying length. They are open from 0.01-0.3 mm, and occur at a density of 60-900 m⁻¹. The maximum concentration of fractures occurs in OM-rich beds of water-sensitive clays, and particularly at the contact of these beds with higher-density interbeds. In beds of pale clay, in interbeds of siltstone, dolomite, and marl, the frequency of fractures successively decreases, until they are completely absent. For this reason, productive units of bituminous clays contain alternately permeable (0.178·10⁻¹² m²), and slightly to impermeable (0.004·10⁻¹² m²) beds.

Researchers hold divergent opinions as to the origin of fractures in clays. They appear to be related to irregular tangential stresses within rocks (the Bazhenovsk Formation of Western Siberia) during the geological past, and at present [29]; to paleoseismic events [22]; to fracture zones [18, 30]; to diagenetic processes [14, 26]; and to the foliation of rocks under the influence of anomalously high pore pressure [14]. Gurari and his successors record that fractures within bituminous clays may be formed, among other causes, by fluid-induced fracturing (oil-, gas-, and autofracturing) in slightly bedded rocks [4, 10, 12, 25, 27]. The process of fracture formation in clays as explained by fluid fracturing of rocks is detailed below.

The oil saturation of the Lower Maikopian sediments demonstrates that they at once form the oil source, reservoir and trap [2, 5, 12]. This combination demonstrates that the oil accumulation was formed in a closed thermodynamic system.

In quasi-isolated systems, the thermal expansion of liquid and gaseous products of OM maturation leads to an increase in fluid pressure [4, 10, 12]. Beyond that, the behavior of the fluid is controlled by a critical pressure value (P_{cr}), equal to the pressure of fluid fracturing of the rock (P_f) [4, 10]. This is defined by the equation:

$$P_f = \frac{0.01 \cdot \nu}{1 - \nu} H \sigma_{bu}$$

where ν is the Poisson's ratio, H is the depth of burial, and σ_{bu} is the bulk density of the overlying rock mass.

The initial pressure of fluid in the fracture corresponds to the equilibrium state of the system (point A on Fig. 1b; Fig. 2a). Line AB corresponds to the processes of compaction and organomineralic transformation of confined argillaceous rocks defined by Le Chatelier's principle. In view of the failure of the OM maturation products to be expelled from the source clays, they remain undercompacted, with densities varying from 2.02-2.18 g/cm³, and porosities of 7.7-17.4%.

With increasing fluid pressure, and its approach towards P_{cr} , the entropy of the system increases. The margin of the hydrocarbon within the fracture (the oil-water-rock interface) becomes unstable, acting as a "weak spot," and becomes mobile (Fig. 2b). As a result, each separate layer of water extends along the fracture, increasing its thickness (as an "aqueous wedge"), and forces apart the fracture margins. Small gas bubbles form in the liquid near the oil-water interface, the presence of which reduces its tensile strength [19].

TABLE 1. Physical Rock Properties and OM Catagenesis of the Lower Maikopian Deposits (Vorob'evsk Region)

Age	Depth, m	Formation pressure, (MPa)	Formation temperature, °C	Rock	Density, g/cm ³	Porosity, %	Catagenetic stage		Productivity
							OM (after Chechua)	clay (after Rassel)	
$P_3^1 - N_1^1$	1500-700	15-7	105-63	Clays	1,96-2,02 1,98 (41)	25,0-27,4 26,6 (41)	$PC_2^1 - PC_1^1$	$PC_2^2 - PC_1^1$	Gas
				Limestone	2,04-2,20 2,12 (9)	23,8-25,2 24,5 (9)			
				Siltstone	1,59-2,08 1,9 (27)	23,7-40,0 30,5 (27)			
$P_3^1 - ^2$	2155-2105	31,5-29,6	135-123	Clays	2,12-2,60 2,41 (139)	2,0-17,7 11,69 (139)	$MC_3^1 - PC_2^2$	$MC_5^2 - PC_3^1$	Oil
				(claystone)					
				Limestone	2,37-2,86 2,70 (73)	0,9-15,4 3,98 (73)			
				Siltstone	2,11-2,56 2,43 (18)	5,6-17,5 11,5 (18)			
K_1	3800-3000	38-30	175-161	Clays	2,25-2,64 2,54 (55)	1,7-16,5 5,12 (55)	$MC_3^2 - MC_1^1$	$MC_5^2 - MC_1^2$	Gas
				(claystone)					
				Limestone	2,31-2,61 2,51 (16)	0,9-3,9 2,87 (16)			
				Siltstone	1,96-2,49 2,22 (81)	2,6-26,9 16,92 (81)			
C_3	> 3800	> 38	> 175	Clays	2,63-2,77 2,71 (28)	0,2-4,3 1,37 (28)	AC_4^{3-6}		
				Limestone	2,61-2,87 2,79 (7)	0,6-2,1 1,3 (7)			
				Siltstone	2,58-2,71 2,66 (19)	0,3-8,00 2,94 (19)			

Note. The numerator denotes the significance limits of the parameter, and the denominator the average value, with the number of analyses in brackets.

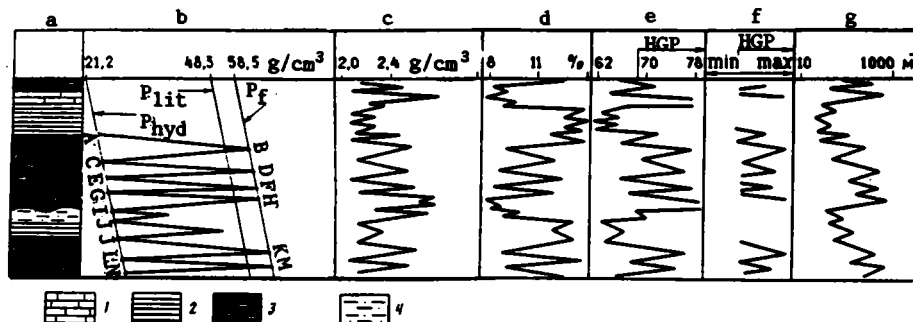


Fig. 1. Theoretical model for the formation of reservoir porosity in Lower Maikopian clays. a) Lithological log; b) pressure; c) density; d) porosity; e) OM maturity, $10 R_v$; f) amount of hydrocarbon generation; g) fracture density. 1) Dolomite (limestone, marl); 2) grey, non-bituminous clays; 3) black bituminous clays with fractures; 4) siltstone. P_{lit} , P_{hyd} , and P_f correspond to the lithostatic, hydrostatic and fluid-fracture (critical) pressure respectively. HGP corresponds to zones of hydrocarbon-generating potential for the relevant parameter.

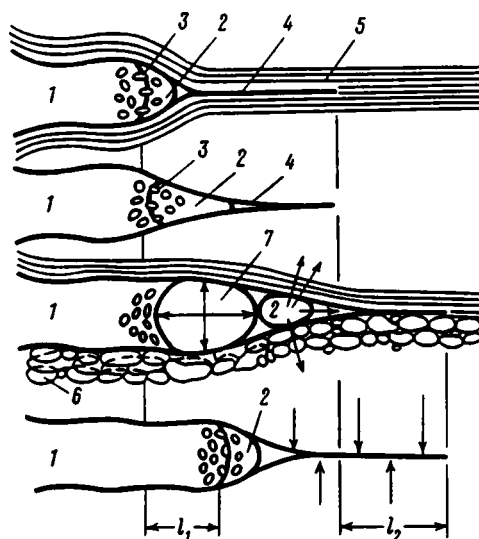


Fig. 2. Mechanism of formation and growth of fractures. 1) Oil; 2) water; 3) gas bubbles at the oil-water interface; 4) fracture margins; 5) clay laminae; 6) siltstone bed; 7) cavitation bubble (cavity). l_1 and l_2 are the lengths by which the oil and the fracture respectively have extended along the lamination.

The critical pressure is reached at point B (Fig. 1b). Once P_{cr} is attained, the sudden lowering of pressure within the fractures causes water to break out from the microstructure of the liquid (Fig. 2c). With the establishment of a pressure gradient, small jets of fluid are forced within $1-2 \cdot 10^{-5}$ sec into the developing fractures, breaking apart the originally intact material (fluid fracturing). The energy of fluid fracturing is so great that in some cases the advancing fracture splits crystals of carbonate and quartz [16]. The pattern of fluid flow in the region of the fracture is very complex and not fixed. This theoretical process is known as the "periodic breakthrough of water" [33]. During one cycle, fluid fracturing may extend a fracture by 5-7 mm (l_2 , Fig. 2) [12]. The growth of formation of embryonic fractures, oriented primarily parallel to bedding, generally follows sedimentary features such as bedding, lithological contacts, fish-scale imprints, etc. The process of stress release in the system is of an explosive character, under the influence of which fractures are propagated at velocities of 0.3-0.7 times that of longitudinal sound waves [12, 16, 21].

Due to the abrupt change of flow velocity in the fracture, a cavitation bubble is formed at the point of fracture propagation (where pressure is instantaneously lowered). The bubble

behaves as a component of the liquid. Following its explosive expansion and attainment of its final dimensions (in about 0.6-0.8 msec), the bubble bursts, producing a hydrodynamic shock wave.

According to the behavior of compressible fluids, the initial pressure in the center of the bursting bubble varies from 57-677 MPa [11, 19]. Once it has burst, shock waves in the compressed media are propagated uniformly in all directions, travelling in excess of the speed of sound [15]. The shock-wave energy manifests as a deformation (compression) of the fracture margins. On their freshly formed surfaces, pressure values may locally reach 220-2500 MPa [15, 19]. Clays with initial densities of 2.0-2.15 g/cm³ may be altered by the shock wave to claystones with average densities of 2.41 g/cm³ (Table 1). These claystones acquire mechanical strength, brittleness, and the ability to retain open fractures. If the fracture walls are of limestone or siltstone, the cavitation may cause irreversible displacement deformation to the adjacent fracture surface. The density of these rocks when affected by the shock wave increases from 2.11-2.37 to 2.56-2.86 g/cm³, and the fracture surface itself is fragmented (Fig. 1c, 2c). This process produces structures in the siltstones which are not typical for such a depth of burial, such as the mechanical compaction of hard clastic grains.

A localized high-temperature (>300°C) pulse is generated on the hard fracture surface at the instant of hydrodynamic shock [1, 34, 35]. In our opinion, this temperature is such that the temperature of fracture formation in clays may reach the transition temperature of marcasite to pyrite, which is 450°C [1]. Thus, wherever pyrite crystals are present on the fracture surface, the local temperature must have reached or exceeded 450°C, whereas the presence of marcasite indicates that temperatures did not reach this value. The redistribution of the heat resulting from the shock mechanism moves in the direction of thermal equilibrium, leading to the formation of a geothermal anomaly in the Lower Maikopian oil-saturated deposits. In wells drilled to a depth of 2100 m within the limits of the anomaly, an increased formation temperature has been measured varying in different parts of the oil field from 123-135°C. These are 20-40°C higher than the regional average for values at the same depth in the Central Precaucasus. In wells which penetrate to basement rocks, the maximum geothermal gradient (6-12°C compared to the regional 3-5°C) is found in oil-saturated Lower Maikopian and Upper Eocene carbonate-claystone deposits, which is an indication of the depth of operation of oil-generating processes.

A strict relationship between the distribution of formation temperatures and productive horizons is not observed from hypsometry. The absence of thermal equilibrium in the deposits shows primarily that fracture formation and the accumulation of hydrocarbons is continuing at present. Their generation does not coincide in time across the whole region, and the hydrocarbons formed at any point in time do not display any distinctive characteristics.

During the mechanical processes of fracture formation, the resulting liberation of energy causes the formation of molecule-sized aggregates. Although compaction reduces the total volume of the rocks, it not only does not prevent, but it actually forces and directs the processes of formation and accumulation of fluid layers in the clays. These processes take place despite the increased temperature of the system and, as is well known, they stimulate the transition from more dense to less dense structures, i.e., in the direction of transformation, which counteracts the effects of increased temperature [1, 19].

The mechanical processes and high temperature acting on the fracture margins increase the reactivity of the rocks, leading to the formation and migration within them of unstable structural defects, and a greatly increased molecular mobility. This results in the appearance of short-lived energetic centers (interior entropy generators), the destruction of which is accompanied by interactions with surrounding molecules. Surplus free energy is utilized in energy-consuming mechanochemical and thermocatalytic processes. The rate at which energy is supplied and absorbed, and also the reaction rates, can be measured in fractions of a second (down to 10⁻⁶ sec). The energy pulse simultaneously leads to a large number of reactions, and a wide range of activation energies (E_a). This energy, necessary for the formation of free radicals, may instantaneously attain and exceed the energy of E_a of the intermolecular bonds. The shock mechanism apparently stimulates several self-supporting reactions on account of an exothermic effect (for example, in the destruction of the lipid-polymer components of OM) [24]. The detailed mechanisms of the compressive shocks, and the kinetic and energetic processes involved are described in [1, 6, 9, 11-13, 15, 16, 19, 24, 28, 31].

The most important result of the mechanical and thermal reactions and processes is the compaction and transformation of organic matter. This initiates the generation of oil, natur-

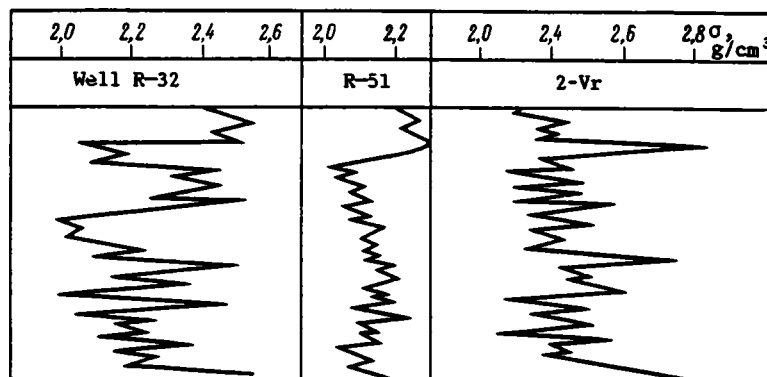


Fig. 3. Bulk density of the Bazhenovsk (wells R-32 and R-51) and Lower Maikopian (well 2-Vr) formations.

al gas, catagenetic fresh water, the direct synthesis and selective reactions of iron with hydrogen sulfide, etc. The geochemical consequences of these processes in the Oligocene argillaceous rocks include the mass recrystallization of carbonates (alteration of calcite to dolomite), or the growth of large dolomite crystals in a fine matrix; the intense hydromicratization of clay minerals; the alteration of marcasite to pyrite, etc. The accelerated maturation of OM (from protocatagenetic grade to MC_3^1) and diagenesis of argillaceous rock (from PC_3^1 to MC_5^2) over an unusually short depth interval (20-60 m) should particularly be noted. Instead of the expected regular increase in organic matter maturity with depth, alternating low and high vitrinite reflectance (R_V) values are often observed. Maximum R_V values are observed in overcompacted zones (Fig. 1e). We have called these "catagenetic anomalies." The R_V value corresponding to the MC_1^1 grade is reached at only 1950 m. By comparison, in the Maikopian clays of the Western Precaucasus (the Indolo-Kubansk Trough) the boundary between the PC_3^2 and MC_1^1 catagenetic grades, with an average temperature of 138°C, occurs at a depth of 3.5 km [17].

The impact stress during fluid fracturing and cavitation processes is localized on the fracture margins, and major dissipation of mechanical energy into the deep rock matrix does not occur. This explains the frequent alternation in oil-saturated parts of the section of overcompacted and relatively undercompacted rocks with very distinctive physico-chemical properties. One of the causes of undercompaction in rocks, displacement deformation at the front of a shock wave, has been demonstrated experimentally [13, 16]. It is known both on a theoretical basis and from experiments that undercompaction is also caused by the hydromicratization of montmorillonite, the dolomitization of carbonates, and the reduction of OM. Overcompaction by fraction formation in bituminous clays is accompanied by the appearance along fracture walls of conformable neoformed organomineralic aggregates with anomalous density. We consider these to be concretions and crystals of iron sulfide, small carbonate concretions from 2-5 mm across equidimensional secondary dolomite grains ($3 \cdot 10^{-5}$ mm), and also pods and lenses of hard bitumen with sizes from $0.4-1.1 \cdot 10^{-3}$ mm. As a result, processes of overconsolidation dominate in the system, leading to an average increase in density and decrease in porosity of the oil-saturated Lower Maikopian rocks. The fracture density increases in the more compact lithologies (Fig. 1g). It is therefore not coincidental that the most productive zones of the Maikopian and Bazhenovsk Formations correspond to the highest values of background rock density [3, 4, 12, 23, 31]. Within the oil-saturated zone, the density variation is less regular than outside it. Throughout the zone we have studied, the most highly productive intervals occur where the most variable bulk-density values are recorded (e.g., 2.20-2.85 g/cm³ in well 2 of the Vorob'evsk area, and 2.0-2.6 g/cm³ in well R-51 of the Salysk area). In low-productivity and failed wells the variation in density values is slight (e.g., 2.1-2.2 g/cm³ in well R-32 of the Salysk field, Fig. 3) [4, 12]. It should be emphasized that this density anisotropy is observed in lithologically uniform beds only 0.1-0.3 m thick in the Lower Maikopian (2.18-2.54 g/cm³) and the Bazhenovsk (2.21-2.73 g/cm³) formations [23].

The hydrodynamic shocks on the oil in fractures give rise to numerous small gas bubbles, which are embryonic cavitation bubbles (Fig. 2c). Under the influence of the shock, liquid (isolated water) within disconnected fractures is instantaneously introduced into the pore

space of the undercompacted matrix rock, leading to an increase in pore pressure by a factor of between 1.35-1.65. This process produces a temporarily open system (Fig. 2c).

With the introduction of overpressured water into the rock matrix from the closed system, a part of the potential energy of the system is expended. Owing to the sudden pressure decrease, the fracture walls and terminations are closed. The fractures are replenished by the segregation and movement into them of newly formed fluids, which occupy the increased fracture volume (l_1 , Fig. 2d). The hydrocarbons at the fracture terminations are bounded by the latest generation of catagenetic water. The system once more becomes metastable, in a state of static equilibrium, isolated, and with minimum values of free energy and entropy (point C on Fig. 1b).

On sections CDE, EFG, etc., of Fig. 1b, the above process is repeated at a frequency determined by the intensity of organic matter alteration, the generation of hydrocarbons, and formation of fractures. This leads to the existence within the system of two possible states. One state, represented by points A, C, E, etc., is characterized by minimum values of density, degree of OM transformation, and intensity of fluid formation and fracturing of the rock. The other state (points C, D, etc.) corresponds to the maximum values of these parameters (Fig. 1b).

Finally, the following conclusions may be made:

1. Processes which form reservoir porosity and lead to hydrocarbon accumulation in bituminous clays, and which are responsible for the productivity of Lower Maikopian deposits of the Central Precaucasus, produce the following anomalous phenomena: an increase in temperature, formation and pore pressure, and also in the density and catagenic anisotropy of rocks and organic matter.

2. Maximum values of temperature, pressure, and rock density, in addition to the intensity of physicochemical reactions inorganic matter, are reached as a result of the fluid-induced fracturing of the rocks, and the consequent cavitational effects. This intermittent process in bituminous clays leads to an increase in reservoir volume due to the generation and growth of fractures, and also to the increase of oil accumulations of a complex nonanticlinal (catagenic) type.

3. The processes of formation of reservoirs in clays, of oil accumulations, and of the anomalous phenomena, are linked in time, location, and origin.

4. Contrasting rock density values in sections of clay deposits may act as a criterion for recognizing the presence of oil.

LITERATURE CITED

1. E. G. Avvakumov, *Mechanical Methods of Activating Chemical Processes* [in Russian], Nauka, Novosibirsk (1986).
2. L. V. Andrieichenko, A. V. Bochkarev, V. N. Evik, and V. P. Il'chenko, "Exploration prospects for hydrocarbons in Lower Maikopian clays of the Central Precaucasus," in: *Exploration and Reconnaissance for Gas in Mature Gas Provinces* [in Russian], VNIlgaz, Moscow (1984), pp. 80-84.
3. I. I. Bobrovnik and L. G. Mitkov, "Acoustic structures in the Bazhenovsk Formation of Western Siberia in relation to the direct exploration for hydrocarbons," in: *Management of Hydrodynamic Processes in the Production and Exploitation of Oil* [in Russian], Zap-SibNIGNKI, Tyumen' (1986), pp. 35-41.
4. F. Ya. Borkun, "Evolution of migration mechanism and reservoir formation in the Bazhenovsk Formation with increasing burial depth," in: *Hydrocarbon Formation at Great Depth* [in Russian], MGU, Moscow (1986), pp. 358-361.
5. A. V. Bochkarev and V. P. Il'chenko, "Geochemical and hydrogeological conditions for the genesis of oil from organic matter in Maikopian deposits of the Central Precaucasus," in: *Aspects of the Relationship of Oil to Organic Matter in Rocks* [in Russian], Nauka, Moscow (1986), pp. 73-75.
6. A. V. Bochkarev, "Kinetics of catagenetic processes," in: *Geochemical Exploration for Oil and Gas* [in Russian], IGU, Irkutsk (1987), pp. 53-54.
7. I. A. Burlakov, R. T. Nalband'yan and N. V. Gullii, "Physicochemical properties of oil (and gas) in the Khadumsk claystone deposits of the Eastern Precaucasus," *Neftepromysl. Delo Trans. Nefti*, No. 7, 26-28 (1985).

8. I. A. Burlakov, R. A. Khadisova, L. A. Keligrekhashvili, and T. B. Leshchinskaya, "Geochemical characteristics of Oligocene oil-bearing clay deposits in the Eastern Precaucasus," *Geol. Nefti Gaza*, No. 4, 40-43 (1986).
9. P. Yu. Butyagin, "Kinetics and nature of mechanochemical reactions," *Usp. Khim.*, 40, No. 11, 1935-1959 (1971).
10. F. G. Gurari and I. Ya. Gurari, "Formation of hydrocarbon accumulations in the Bazhenovsk Formation of Western Siberia," *Geol. Nefti Gaza*, No. 5, 36-40 (1974).
11. G. V. Draiden, A. P. Dmitriev, Yu. P. Ostrovskii, and M. I. Éttingberg, "Shock waves formed in water by the bursting of cavitation bubbles," *Zh. Tekh. Fiz.*, 53, No. 2, 311-314 (1983).
12. V. N. Evik, and A. V. Bochkarev, "Mechanism of formation of reservoir and oil accumulations in Oligocene black shales of the Central Precaucasus," in: *Geology, Mineralogy and Lithology of Black Shales* [in Russian], KF AN SSSR, Syktyvkar (1987), pp. 115-117.
13. B. V. Zamyshlyayev, L. S. Evterev, S. G. Krivosheev, and Yu. P. Pilipko, "Model for dynamic deformation and destruction of rock masses," *Dokl. Akad. Nauk SSSR*, 293, No. 3, 568-571 (1987).
14. S. G. Zaripov, V. P. Sosnin, and K. S. Yusunov, "The reservoir formation mechanism in the Bazhenovsk Formation," in: *The Oil-Bearing Bazhenovsk Formation of Western Siberia* [in Russian], IGI RGI, Moscow (1980), pp. 62-68.
15. Ya. B. Zel'dovich and Yu. P. Raizer, *Physics of Shock Waves and High-Temperature Hydrodynamic Phenomena* [in Russian], Nauka, Moscow (1966).
16. J. A. Zukas, T. Nicholas, Kh. F. Swift, et al., *Dynamics of Shocks* [Russian translation], Mir, Moscow (1985).
17. S. G. Nerucheva (ed.), *Catagenesis and Hydrocarbons* [in Russian], Nedra, Leningrad (1981).
18. T. T. Klubova, L. P. Klumushina, and A. M. Medvedeva, "Formation of oil in Bazhenovsk Formation clays," in: *The Oil-Bearing Bazhenovsk Formation* [in Russian], IGI RGI, Moscow (1980), pp. 69-84.
19. R. Knapp, J. Daley, and F. Khemmit, *Cavitation* [Russian translation], Mir, Moscow (1974).
20. V. S. Kosarev, "The oil-bearing Lower Maikopian deposits of Eastern Stavropol'," in: *Scientific Principles for Hydrocarbon Exploration* [in Russian], IGI RGI, Moscow (1985), pp. 119-128.
21. V. S. Kuksenko, A. I. Lyashkov, K. I. Mirzoev et al., "Relationship between the size of fractures formed under load, and the duration of elastic energy emission," *Dokl. Akad. Nauk SSSR*, 264, No. 4, 846-848 (1982).
22. K. I. Mikulenko, "Prospective hydrocarbon-bearing deposits of the Bazhenovsk Formation [in Russian], SNIIGGIMS, No. 196, Novosibirsk (1974), 37-41.
23. L. A. Nazina and G. P. Gubina, "The role of petrophysical studies in determining the reservoir capacity of Bazhenovsk Formation rocks," in: *Hydrocarbon Formation at Great Depth* [in Russian], MGU, Moscow (1986), pp. 186-187.
24. S. G. Nerucheva (ed.), *Hydrocarbon Formation in Domanikov-Type Deposits* [in Russian], Nedra, Leningrad (1986).
25. V. I. Petrenko and A. V. Bochkarev, "Hydrodynamic formation of gas from coal," *Sov. Geol.*, No. 12, 35-40 (1987).
26. G. É. Prozorovich, A. P. Sokolovskii, and A. G. Malykh, "New data on fracturing of Bazhenovsk Formation Reservoirs," *ZapSibNIGNI, Probl. Nefti Gaza Tyum.*, No. 18, 7-9 (1973).
27. M. F. Svishchev, M. M. Sadykov, N. D. Kantelin, and K. S. Yusupov, "Hydrodynamic features of productive beds in the Bazhenovsk Formation of the Salym oilfield," in: *Geology and Oil Development in Western Siberia* [in Russian], ZapSibNIGNI, Tyumen' (1973), pp. 239-251.
28. A. A. Trofimuk, N. V. Cherskii, V. S. Vyshemirskii et al., "Natural factors responsible for the transformation of buried organic matter," *Geol. Geof.*, No. 6, 72-77 (1982).
29. A. A. Trofimuk and Yu. P. Karagodin, "The Bazhenovsk Formation: a unique natural oil reservoir," *Geol. Nefti Gaza*, No. 4, 29-33 (1981).
30. É. M. Khalimov and V. S. Melik-Pashaev, "Exploration for economic oil accumulations in the Bazhenovsk Formation," *Geol. Nefti Gaza*, No. 6, 1-10 (1980).
31. V. S. Kharin, "Influence of thermobaric conditions on the formation of the productive zone of the Bazhenovsk Formation," in: *Hydrocarbon Formation at Great Depth* [in Russian], MGU, Moscow (1986), pp. 362-363.
32. G. N. Chepak, V. M. Shaposhnikov, P. S. Naryzhnyi, et al., "Oil-bearing clay bodies in the Oligocene of the Eastern Precaucasus," *Geol. Nefti Gaza*, No. 8, 36-40 (1983).
33. L. A. Épshtein, "Characteristics of ventilated caverns and some effects of scale," in: *Unstable, High-Velocity Water Currents* [in Russian], Nauka, Moscow (1973), 173-185.

34. P. Bankwitz, "Probleme von Bruchentstehung in Bruchablauf in Gesteinen sowie deren Bedeutung in Geologie und Geophysik," FMC-Ser. Inst. Akad. Wiss. DDR., No. 2, 49-63 (1982).
35. I. Taylor, "Cavitation pitting by instantaneous chemical action from impacts," ASME, Paper 54-A-109, 1954.

PALEOGEOGRAPHY OF THE COLCHIS LOWLAND PEAT AREA DURING THE HOLOCENE

L. I. Bogolyubova

UDC 551.8:553.97(479.2)

For the first time, the Holocene paleogeographic evolution of the Colchis Lowland peat area is being described by the method of detailed lithofacies analysis.* Seven paleogeographic charts, covering the last eight thousand years indicate an interrelated pattern of change in the paleolandscape units in time and space. This change is displayed by the character of the configuration and alteration patterns between paragenetically interrelated units. Study results from Colchis Lowland peat deposits, viewed as developing coal formations in an intermontane basin that is open toward the sea, provided conclusive facts for the reconstruction of paleolandscapes and sediment genesis of corresponding ancient coal-bearing formations.

The Holocene paleogeography of the Colchis Lowland that also includes an area of peat accumulation has been discussed in a number of papers [3, 4, 9, 11, 17]. The data are sketchy and stratigraphic information predominates. The papers lack paleogeographic charts and are restricted to discussing the regressive and transgressive Black Sea stages, based on the presence of marine deposits and peat beds in the sections that usually are not illustrated in the papers. They also utilized lithologic studies of separate marine and continental terraces. These articles included pollen-spore data from the deposits, useful in the stratigraphic and paleoclimatologic evaluation of the Holocene and reconstruction of the contemporary plant cover pattern. Archeological finds and absolute ages that appeared in recent years [4] were used in dating peat horizons that marked the beginning of the Black Sea regression. Marine units, associated with the transgressive phase, were similarly utilized.

Our studies of the Colchis Lowland Holocene paleogeography were based on a network of lithofacies profiles, newly compiled [2]. These, along with absolute age dates along many sections provide the required control in constructing a set of detailed paleogeographic charts (nine in the Poti Basin; four in the Kobuleti Basin) for each one thousand year-time interval throughout the Holocene.

This study's data base is represented by 600 drillholes; 10-25 m deep, drilled by us manually or power-driven equipment. About 100 drillholes penetrated 250-300 m deep; the core material was obtained by hydrogeological field parties. For few drillholes descriptions, columnar logs and sections were obtained from geological reports and publications [8, 10, 14, 17, 18]. The material was analyzed by the lithofacies method, employed for the first time in the subject area, as discussed in detail [2]. The paleogeographic charts were compiled on the basis of an earlier developed method, used in coal-bearing formations [5, 15]. Without details, only the essentials only are presently explained. The approximate outlines of the landscape units are displayed on the maps. These define the physical conditions and settings of sedimentation, in turn defining the paleogeographic conditions of sediment and peat deposition in the Colchis Lowland during a given evolutionary stage. Definition of the landscape unit contours and paleogeographic reconstruction indicate the distribution of shoreline positions, the direction of the stream flow, location of and interrelationship between stream channels, floodplains, lakes, swamps, vegetatively overgrown lakes and lakes that became converted into swamps, etc. These data result from the synthesis of our assumptions, based on the described facies and their combinations that existed in given time intervals

*A detailed geological description of the peat area was provided by a previous paper [2].

Geological Institute, Academy of Sciences of the USSR, Moscow. Translated from *Litologiya i Poleznye Iskopaemye*, No. 1, pp. 69-90, January-February, 1990. Original article submitted January 5, 1989.

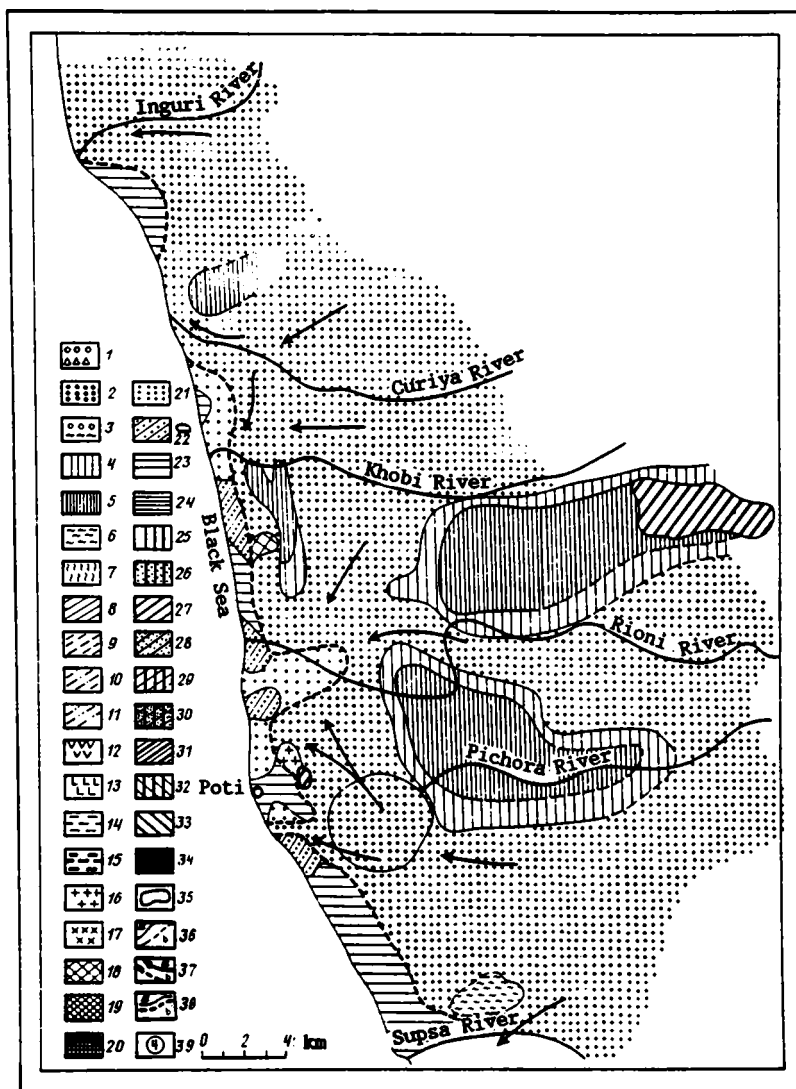


Fig. 1. Schematic paleogeographic map of the Colchis Lowland peat area of the Poti Basin c. 7-8,000 yr. B.P. Paleolandscape elements: 1) river stream channel (ARG); 2) lowland river channel (ARP); 3) alluvial fans of mountain streams, piedmont areas, including valley trains (PPK); 4) river floodplain (APP); 5) interior floodplain (APV); 6) lakes in forested floodplain areas (APZ); 7) lakes in the nonforested floodplain areas (APO); 8) oxbow lakes (APS); 9) sapropelic fresh water lakes (OSP); 10) sapropelic lakes with traces of salt water intrusion (OSS); 11) highly brackish sapropelic lakes (OSO); 12) lake bodies, overgrown by vegetation and mud-filled woody-grassy peat swamps (OZD); 13) overgrown lake bodies and mud-filled grassy peat swamps (OZG); 14) forested river valley areas, converted to swamps (woody soils-PDD); 15) swampy grassland areas of stream valleys (grassy soils-PDT); 16) swampy forested coastal areas (woody soils-PMD); 17) swampy coastal grassland areas (grassy soils; PMT); 18) very low energy marine bay areas (MPZ); 19) very low energy lagoonal areas (MPL); 20) coastal barrier ridges and beaches (MMV); 21) coastal deltas (MMD); 22) high energy shallow marine basin areas (MMK); 23) wave-impacted shallow marine areas (MMM); 24) delta margins and relatively distant marine basin areas (MUM). Lower peat swamps; 25, 26) woody; stagnant, respectively, with water circulation; 27, 28) grassy; stagnant, respectively, with water circulation; 29, 30) grassy-woody; stagnant, respectively, with water circulation. Transitional peat swamps: 33) *Imbricatum*-moss limit of peat accumulation; 36) landscape boundary (a, factual; b, assumed); 37) stream flow direction (a, factual; b, assumed); 38) shortline (a, present; b, ancient); 39) peat deposits: 3, Zortatsk; 4, Anakliisk; 5, Churiisk, 6) Nabadsk, 7, Poti; 8, Pichorsk; 9, Palimastomsk; 10, Imnatsk; 11, Moltakvsk.

in the area of peat and sediment accumulation. Our definition of the term "facies," in agreement with Zhemchuzhnikov [7] and Timofeev [15] refers to sediments with complex of genetic characteristics, formed under conditions, defined by a given landscape unit (sediments and depositional conditions). Compilation of paleogeographic charts, based on highly detailed facies analysis data, was this study's specific objective. A very short geologic time interval, only ten thousand years, was covered within a small areal.

Opinions vary in the USSR and abroad on the position of the lower Holocene boundary. In the last decades, as the result of a steady stream of radiocarbon data, the researchers have practically reached agreement in fixing the boundary between 10-12 ka B.P. The question is avoided here. In the Colchis Lowland the start of the Old Black Sea transgression is accepted for the lower Holocene time boundary. It occurs directly beneath sandy-aleurolitic-clayey marine deposits that in the Colchis Lowland are sharply separated from the underlying Novo-Euxinic sediments. The Novo-Euxinic deposits in Drillhole #13 (on the right bank of the Rioni River, 13 km from the Black Sea) are composed of mottled grayish brown sands, aleurolites, and clays with iron-enriched patches and veins of siderite spherulites, layers of buried soil remnants of roots systems, replaced by Fe-hydroxide and occasionally by authigenic montmorillonite. These have subaerial origins. In the Kobuleti Basin, the Novo-Euxinic deposits [2] are represented by bouldery-gravelly mountain stream sediments, overlain, as shown by us [2] and Tvalchrelidze [14], by Old Black Sea continental deposits. The boundary between Old Black Sea and Novo-Euxinic deposits [2], occurs in various Poti and Kobuleti Basin drillholes at different depths. Absolute dates from the plant matter of the boundary layer at Drillhole #S-26 (mouth of the Khobi River, Kulevi area) at 38 m depth produced an age of $10,550 \pm 200$ yr B.P. (GIN-632). This establishes the start of the Old Black Sea transgression and, consequently, the start of Holocene sedimentation in the Poti Basin. According to Balabanov and others [1] (second phase of the Black Sea transgression), the highest layer in the Novo-Euxinic sands and clays on Pitsunda Peninsula is dated 9960 ± 340 yr. B.P. (Li-369). This value is close to the date, obtained near the mouth of Khobi River. In summarizing research data on the basis of absolute ages in the area of Sochi, Adler, the Azov Sea bottom, Pitsunda Peninsula, the mouths of the Khobi and Supsa Rivers, Dzhanelidze [4] concluded that the final phase of the last Novo-Euxinic Black Sea stage occurred c. 10-10.5 ka B.P. According to him, these data are in agreement with conclusions of Degens and Hunt who dated upper Novo-Euxinic beds in the Black Sea 9-10 ka B.P. These data confirmed the age of the beginning of the Holocene, established by A. A. Vinogradov and S. M. Devirtsem and others in the Bosfor region as 9350 ± 220 yr. B.P. (Mo-287).

Three main factors control Holocene and modern sedimentation in the Colchis Lowland: (1) movements of tectonic uplift in the Greater and Lesser Caucasus ranges; (2) deltaic sedimentation of rivers that originated in these ranges; (3) eustatic rise of Black Sea level. These processes changed the landscapes within the Holocene framework of the Colchis Lowland paleogeography, resulting in more or less complex combinations of the genetic categories and their facies patterns. Such patterns can easily be demonstrated on a series of paleogeographic charts.

The map that covers the Old Black Sea period (start of Holocene, c.10 ka B.P.) indicates that sea waters covered a large area in the western Colchis Lowland of the Poti Basin.* In the present land area the sea extended farthest in the Rioni River Valley, c. 14-16 km from the present Black Sea shoreline. Further north- and southward (toward the basin rim) its distribution zone narrows. Approximately the same Old Black Sea shoreline of the Old Black Sea shoreline is shown also on Tsereteli's paleogeographic chart in the Poti Basin [17]. East of the old shoreline undifferentiated deposits of lowland streams and, partially, mountain floodplain streams occupy the rest of the Sea deposits in the Kobuleti Basin are almost everywhere represented by glacial meltwater, diamicton (mixed, stony-muddy) deposits [2]. A very narrow zone north of the Choloki and Natanebi Rivers, near Kobuleti, covered by the sea at that time, is an exception [13, 14]. Consequently, at the start of the Holocene the Colchis Lowland in the Kobuleti Basin was almost entirely a land area. The continental facies of the Old Black Sea stage in the Kobuleti Basin were correctly characterized by Tvalchrelidze [13, 14], as having an excess of terrigenous matter, compensating for the fast rate of glacio-eustatic sea level rise and the tectonic subsidence of land areas [3, 4]. The large sediment

*In the Abkhaz-Mergel'sk Basin a significant part of the Colchis Lowland and its peat area is covered by Black Sea waters. These areas form the sea floor. The paleogeography of this area, therefore is not discussed here.

volume may be explained by intensive melting of the Valdai Glacier at the start of the Holocene on the nearby Adzhar-Trialetsk Mts. Evidence of this glaciation (cirques, kars, glacial valleys) was noted by Tvalchrelidze [13] along the ridge line (Khino, Taginauri and Bakhmaro areas) and in the active uplift of these mountains. This uplift produced additional alluvial matter for the streams.

In a number of Poti Basin drillholes that encountered units of the Old Black Sea stage, peaty clay and peat beds were also encountered among the other deposits. These generally are thin (0.1-0.4 m; occasionally thicker) and occur at 36-37 m depth. The beds were considered by Tsereteli [17] as corresponding to the "Colchis Regressive Phase" of the Black Sea. However, the deposits lack the required evidence for a detailed regional lithofacies study. Therefore, we can provide only a generalized summary in the paleogeographic chart of this time interval.

According to our data [2], the New Black Sea sediment- and peat-deposition cycle followed accumulation of the Old Black Sea stage deposits in the Colchis Lowland (Poti and Kobuleti Basins). Its lower boundary in the Poti Basin is marked by lagoonal clay, enriched in humus-bearing attritus (Drillhole #23) or peat beds at 18-20 m depths, occasionally deeper. Their remnants are shown on the map (Fig. 1), in reference to interior floodplains and channel-margin floodplains on the far eastern side of their distribution area. The existence of this horizon had been noted by all workers. However, disagreement surfaced with regard to the chronology of events. Tsereteli [17] attributed the horizon to the regressive "Poti" Black Sea phase, dating it 5 ka B.P. The lagoonal clay, enriched in humus-rich attritus (according to our data, ash-rich peat) was radiocarbon dated 7900 ± 60 yr. B.P. (tb-86). The differences in the chronological subdivisions by various authors were illustrated in detail by Dzhanelidze [4]. Such disagreements result from correlation problems in existing stratigraphic schemes in the Colchis Lowland area, aggravated in addition by the lack of lithofacies analysis that should have been carried out in specific sections.

The start of the New Black Sea cycle in the Poti Basin, covering the time range between 7-8 ka B.P., is shown in Fig. 1. Despite continuing Black Sea transgression, the chart shows a significant reduction in the area covered by the sea. The sea retreated behind the invading flood of stream sediments in zones of prograding stream delta plains. An undifferentiated alluvial sequence formed, the result of "pseudoregression." The intensive sediment supply (20 million $\text{Cu m}^3/\text{yr}$) was due to uplift of the Greater and Lesser Caucasus ranges and relative subsidence of the Colchis Lowland surface [17]. The sea existed in the far west, extending its tongues into continental areas, especially in areas regions subaqueous deltas had been actively forming. The marine deposits formed a narrow, uneven zone off the entire shore of the Poti Basin. In the central part of the high energy shallow marine basin areas sandy deposits (MMK) concentrated (underwater bars, shoals, ridges) and underwater deltas (MMD) that received most of the river sedimentation (Table 1). To the north and south, sandy, high wave energy accumulation occurred on the coastal plains. Prograding delta plains of ancient streams and their channels formed, although the ancient Rioni channel was identified only along a short coastal plain segment. Sandy-silty-clayey floodplain deposits formed (APP and APV) that are almost completely absent in the rest of the area or occupy but small areas.

At the start of the New Black Sea cycle sediment and peat accumulation was characterized by marine intrusion over the continent in the Kobuleti Basin, south of the Colchis Lowland. As the result, the coastal strip of the Old Black Sea continental area became covered by marine deposits. The facies combinations consist of sediments formed in shallow, high wave energy facies (MMM) and underwater deltas (MMD) in the area of the Natanebi and Chopoki Rivers. In contrast with the floodplain alluvium of the Poti Basin, a large part of the land area was occupied by undifferentiated. The mountain alluvium displays individual zones of near-channel and internal floodplain sediments and, in the southern half of the Basin, piedmont-type alluvial fans. There planktonic diatomitic algae (Melosira) indicate its relatively great depth (facies OSP). In the Kobuleti Basin, the continental facies complex therefore at this time generally was represented by coastal mountain valley landscape units, including also a developing sapropelitic lake.

6 ka B.P. (Fig. 2), a new stage started in the development history of the Colchis Lowland in the Poti Basin. The sea continued to recede. Small tongues of marine deposits became preserved in the far north, in the mouth of the ancestral Inguri River and in the southern half of the Basin, especially around the town of Poti. This was an area where these sediments are most deeply buried and where deltaic sedimentation did not interfere with the marine

TABLE 1. Basic Genetic Types, Macrofacies and Depositional Facies of Holocene Colchis Lowland Sediments

Genetic disposi- tional group	Macrofacies	Facies	Index
1	2	3	4
Proluvial (P)	Intermittant stream deposits (9PP)	Gravelly-sandy and silty-clayey, poorly sorted alluvial fan deposits of mountain streams, piedmont and related alluvial train deposits	PPK
Alluvial (A)	River-channel deposits (AP)	Boudary-gravelly mountain river channel and related alluvial fan deposits	ARG
		Gravelly-sandy river mouth deposits and related channel sands of lowland streams (downstream zones)	ARP
	Floodplain deposits (AP)	Sandy-silty floodplain deposits along rive channels	APP
		Sandy-silty-clayey deposits of interior floodplain zones	APV
		Silty-clayey lake deposits in forested flooplain areas	APZ
		Clayey deposits of lakes in open (nonforested) floodplain zones	APO
		Silty-clayey oxbow lake deposits	
Lake basins (O)	Sapropelic lake deposits (OS)	Silty-clayey deposits of sapropelic freshwater lakes	APS
		Silty-clayey deposits of sapropelic freshwater lakes with traces of brackish influence	OSP
		Silty-clayey deposits of sapropelic lakes with minor brackish influence	OCC
		Silty-clayey deposits of overgrown lakes of mud-filled woody peat swamps	OCO
		Silty-clayey deposits of lakes overgrown by vegetation and mud-filled grassy peat swamps	OZD
Swampy nearshore coastal plains of soily formations (P)	Swampy deposits of stream valley areas (PD)	Silty-clayey deposits of swampy, forested stream valley areas	OZT
		Clayey deposits of swampy meadowland areas in stream valleys	PDD
	Deposits of swampy sapropelic lake basin (DS)	Clayey-deposits of swampy sapropelic lake basins, overgrown by vegetation	PDT
	Deposits of swampy coastal zone areas (PM)	Sandy deposits of swampy, forested coastal zone areas	PSD
		Sandy deposits of swampy meadowland areas of the coastal zone	PMD
Coastal marine (M)	Bay and lagoon deposits (MP)	Silty-clayey deposits of low-energy bay areas	PMT
		Silty-clayey deposits of low-energy lagoonal deposits	MPZ
		Silty-sandy shore deposits of small stream deltas, bays and lagoons	MPR
		Sandy deposits of coastline ares lagoons and bays, adjacent to the seashore	MPP
		MMV	MMV
	Nearshore, high energy marine deposits (MM)	Sandy deposits of detals along the seashore	MMD
		Sandy deposits, accumulated in high0-energy shallow marine areas (submarine bars, ridges, shoals, etc.)	MMK
		Sandy deposits of high wave energy shallow, gently-slopping off shore bottom	MMM
	Relatively near marine deposits (MU)	Silty-clayey sediments of detla rims and relatively distant shallow marine basin areas	MUM

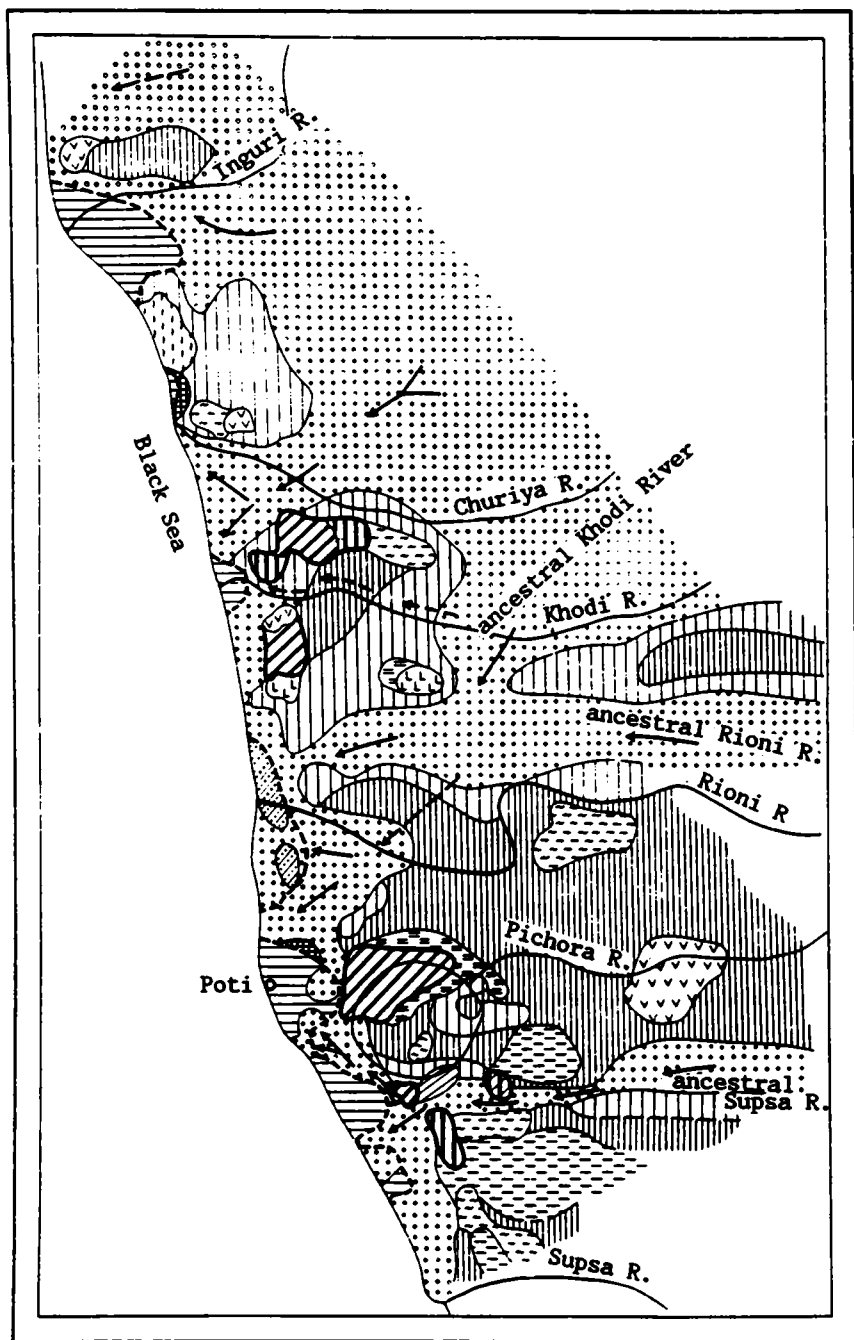


Fig. 2. Schematic paleogeographic map of peat accumulation zone, Colchis Lowland of the Poti Basin, at 6 ka BP. Symbols: Fig. 1.

transgression. From the rest of the Poti Basin (northern and southern half) the sea receded almost completely as the and prograded due to continued, large-scale deposition of alluvial matter. High wave-energy shallow water marine deposits (MMM) predominate in the marine facies group. Small quantities of sand bar and delta (MMD) deposits also occur adjacent to the land area.

The alluvium became differentiated into channel and floodplain types. Wide stream valleys formed; those of the ancestral Inguri, Churiya and Rioni Rivers. The lower course of the ancestral Rioni became combined with the ancestral Khobi Valley. The channel of the ancestral Supsa River is sharply defined. It reached the sea SW of Paliastomi Lagoon. Floodplain deposits that always occur in the interfluvial became broader than they had been in the previous millenium. A large number of floodplain lakes (APO and APZ facies) occurred on

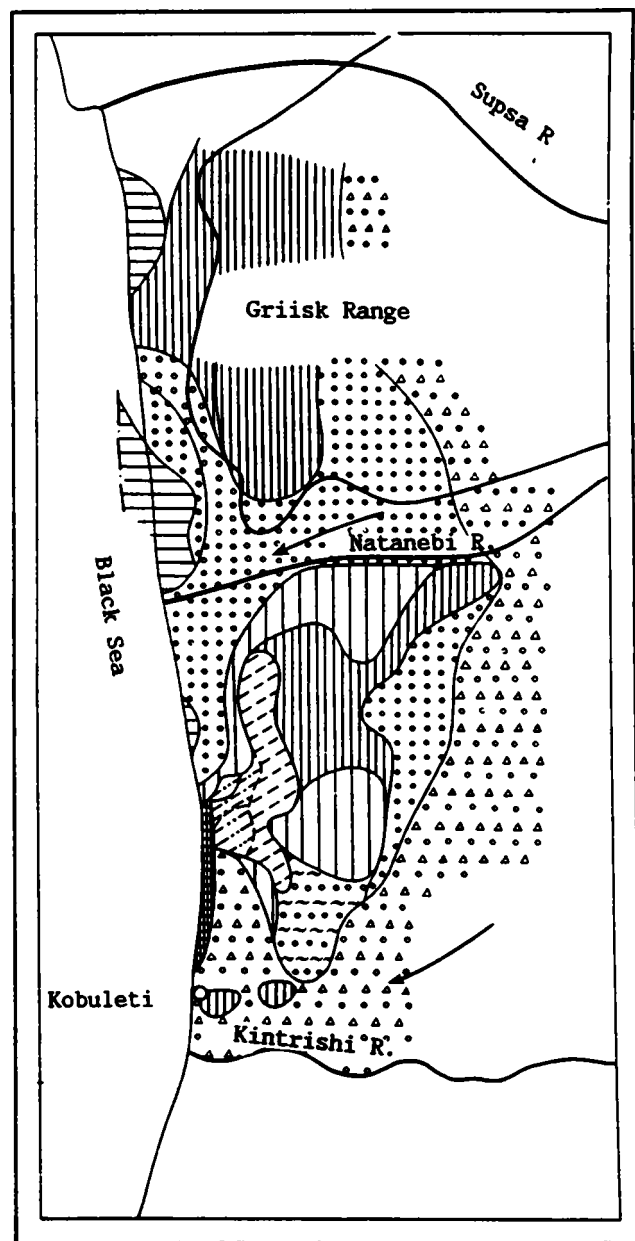


Fig. 3. Schematic paleogeographic map of the Colchis Lowland, Kobuletsk Basin, at c.6 ka B.P. Symbols: Fig. 1.

them, occasionally overgrown by grasses and woody vegetation that form soils of acies PDD and PDT. These units are best displayed on the right and left banks of the ancestral Supsa River. The first indications of peat accumulation appeared in the central parts of the future Nabadsk and Paliastomi peat deposits (Fig. 2) in the form of small, irregular grassy swamp rim, occasionally covered by woody matter. In the Moltakvsk deposit, their cover is essentially woody. The start of peat swamp development was carbon-dated as having occurred between 5950 ± 100 (GIN-1085) and 6480 ± 90 (GIN-1575); occasionally more than 6930 ± 80 yr. B.P. The landscape complex in the Colchis Lowland of the Poti Basin is represented by coastal plain-type alluvial accumulation but with an incipient zone of peat development.

As in the Poti Basin, the landscape units of the Colchis Lowland in the Kobuleti Basin have undergone certain changes at the start of the millenium (Fig. 3). A change from a mountain valley setting to a floodplain-valley type occurred, except in the far south where mountain alluvium (ARC) continued to form along that of the adjacent alluvial train (PPK) to the north. It also accumulated in a narrow east of the subject area. At the same time, interior

floodplain (APV) sediments were deposited in the northern half of the lowland during the change from mountain alluvium deposition. Their distribution area was expanded in the Southern half, in conjunction with channel-rim floodplain (APP) sedimentation. The combined Natanebi and Chopoki River valleys started to take shape, as the result. Underwater delta facies were preserved in river mouth areas. Small portions of interior floodplain deposits (APV) formed north of Kintrishi River among the mountain alluvium. The former sapropelic lake increased its size through the development north- and southward of narrow, smooth-shaped tongue-like projections. The lake's shape and size has therefore changed. A larger, shallow water body existed here, as indicated by the epiphytic bottom-dwelling diatoms that predominated in its sediments. In its western, seaward side the lake was slightly brackish, resulting in sediment facies OSS and OSO, with minor amounts of saline facies diatoms among the predominant fresh water types. At this time, the sea almost completely receded as the result of the prograding shore zone. The abundant sediment volume for this came from lowland and mountain streams in the southern half of the basin. The former marine area was replaced by floodplain channel alluvium and only a small portion of the sea remained, represented by a shoreline indentation in the basin center. Further to the south, in the area of the sapropelic lake, a narrow barrier ridge (MMV) separated land from the sea. Tvalchrelidze [14] did not identify the barrier and mistakenly believed the sapropelic fresh water lake deposits to be part of transgressive marine sediments. In the northern half of the Colchis Lowland in the Kobuleti Basin, the sea remained within its previous limits and, locally in combination with deltaic deposits, formed facies MMM.

The earliest evolutionary stage of landscape units 6 ka B.P. in the Colchis Lowland, Poti Basin, was followed by later development. As the result, the facies combinations shown on the paleogeographic map of 5 ka B.P. are partially similar to the previous map. The main difference lies in the morphology and dimension of the distances involved. The tongue-shaped landward extensions of the sea were reduced everywhere, including at Poti. These tongues are absent in the north, south of the Inguri River. Their shape became irregular as the distribution area was reduced in comparison with the facies area shown on the previous chart. Narrow barrier ridges occur in the northern half of the Lowland. Floodplain areas increased significantly, especially in the ancestral Inguri and Rioni River areas, accompanied locally by channels (APP) by finer, clayey (facies APV) sediments that characterize the bottom of interior interfluvies. At the same time, swamp areas, including adjacent floodplain zones, became wider. The outlines of these fragmented swamp areas, as before, were of irregular shape. In the region of the Nabadsk deposit these areas increased in size. They occupied the base of grassy peat associations that along their marginal zones, locally alternate with woody plant associations. They are represented by corresponding peat and soil types. The Churiya peat swamp area occurs further north. Incipient, small, grassy, peat swamp lenses have been preserved among floodplain and lake (APO) facies in the interfluvial area of the ancestral Inguri and Churiya Rivers (earliest stages of the Anakliisk peat deposit). Peat and soil deposits occur along their northern and southern margins. Individual small peat swamp lenses with grassy and woody peat deposits occurred in the eastern part of the floodplain area in the ancestral Rioni/ancestral Supsa interfluvial area. The facies combination also included small lakes in the forested floodplain area (APZ facies); also overgrown floodplain (PDD) facies, in contact with them or, occasionally, in isolation. The western half of this area was occupied by a quite large peat swamp that represented the early stage of the Poti, Paliastomi, and Moltakvsk peat deposit development. It was underlain by vegetated internal floodplain deposits (facies PDD). They occupied almost the entire territory of the future Paliastomi Lagoon and additional areas; part of the area of the town of Poti, later almost exclusively covered by grassy plant communities. These deposits extended also far south along the basin rim where essentially woody deposits formed the peat body, inherited from the former swamp cover in the Moltakvsk peat deposit area. Thus, the landscape of the Colchis Lowland acquired the appearance of an alluvial-swamp plain with intensive peat accumulation. Widening of the area of floodplain deposit and of the associated peat swamps narrowed the ancestral river valleys. Their channels became more distinct. They preserved essentially the former deposits, including that of the ancestral Khobi River that became the right-hand tributary of the ancestral Rioni River.

At c. 5 ka B.P. no important changes occurred in the general configuration of landscape units in the Colchis Lowland of the Kobuleti Basin. The only change was increase in the diatom content of the acidic fresh water sapropelic lake. At the same time, vegetation covered the lake from a direction of the west where a woody peat swamp started to form. The lake became swampy (facies PSD). The sea continued to recede in the region of the ancestral Natanebi Riv-

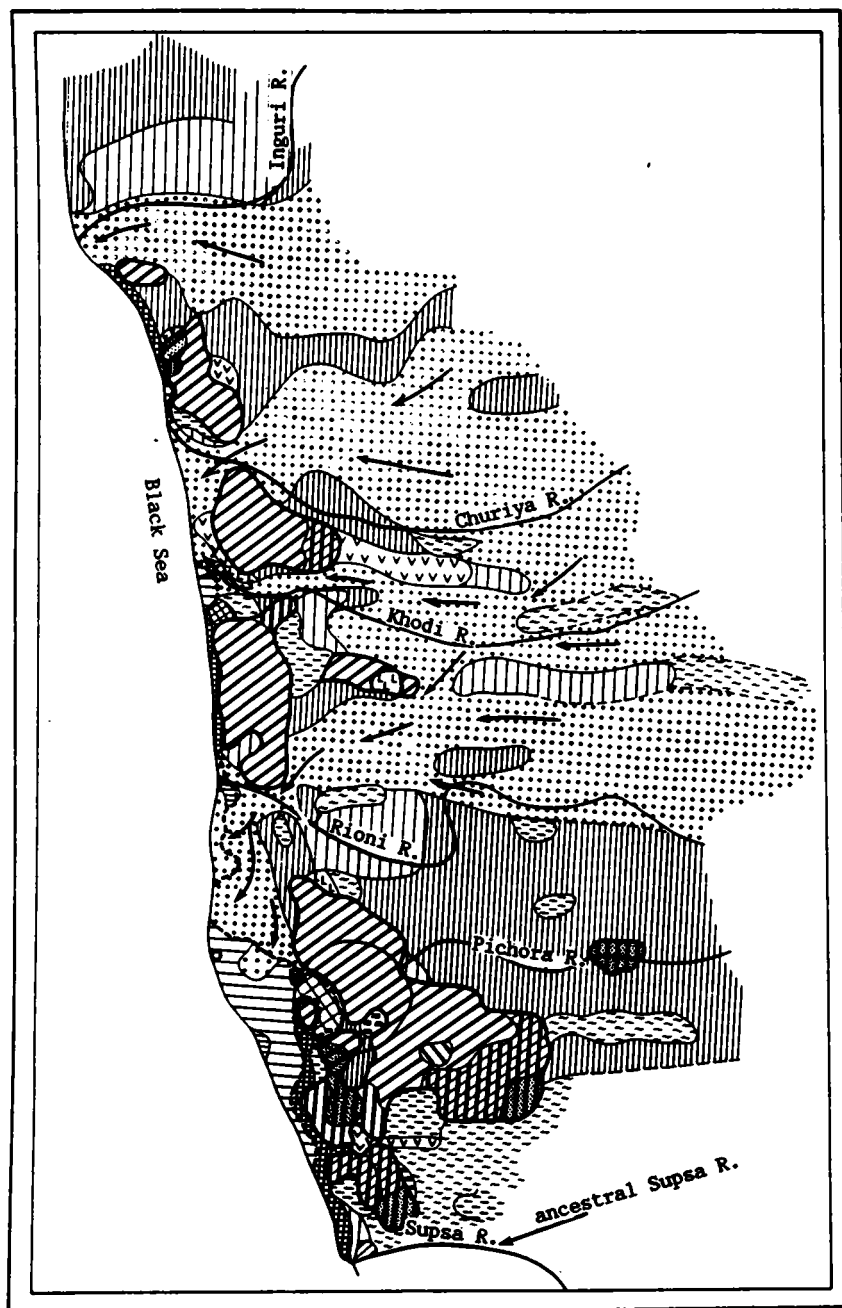


Fig. 4. Schematic paleogeographic map of the peat area in the Colchis Lowland, Poti Basin, at c. 4 ka B.P. Symbols: Fig. 1.

er and north of it. As before the river transported large volumes of continental deposits, forming an underwater delta (MMD). The area of southern floodplain deposits increased somewhat as they occupied new areas to the north.

The main event of the 4 ka B.P. period (Fig. 4) in the Colchis Lowland of the Poti Basin was a marine transgression, clearly evident only in the southern area. The present coastal zone here was occupied by shallow, high-energy (MMM) marine deposits, and certain amounts of relatively deeper water sediments along delta margins and in interdelta areas (MUM). The sea also intruded from the west over the Paliastomi peat swamp, forming a brackish embayment with sapropelic clay deposits (facies MPZ). These contained brackish-marine diatoms and horizons of local molluscan shell concentrates. The absolute age derived from the overlying peat bed were 3930 ± 200 (GIN-1234) and 4580 ± 50 (GIN-1994) yrs. B.P. These ages coincided with the maximum extent of the New Black Sea transgression [2, 4], as established by Fedorov [16]

for other Black Sea areas. The transgressing sea covered this part of the Colchis Lowland entirely up to an elevation of 2 m [16]. This happened in conjunction with global sealevel effects that had several causes. The prime reason was subsidence of the coastal land area, caused by low-amplitude differential movements, typical of individual structural blocks in the Colchis Lowland at this time. This occurred against a background of continued subsidence and subsidence of the entire Rioni intermontane basin in general. Apparently, this process shifted the ancestral Supsa River southward into a more stable position within the given geotectonic setting. The old channel became a large floodplain lake, accompanied by oxbow lakes. West of the lake, the previously noted woody peat swamp developed and its old channel became occupied by the transgressing Black Sea waters. Terrigenous matter from the new channel of the ancestral Supsa River was distributed along the coast as far as the SW coast of the developing Paliastomi Bay. This is how coastal barrier deposits of facies MMV formed. Coastal barrier deposits along the northwestern margin of the bay formed from sediments delivered by the ancestral Rioni River.

As the chart shows, no changes occurred within the rest of the landscape units. Only the distribution areas of all the peat swamps did increase, as they shifted primarily eastward. South of the ancestral Inguri River a small peat swamp formed at the future area of the Tikorsk peat deposit. The swamp attained its previous dimensions in the Anakli area and developed partially through encroachment into a lake, partially through floodplain enlargement. As before, the central part of the peat swamp was constantly occupied by a grassy plant community, rimmed by a woody plant community, also participating in the overgrowth of the lake. The formation processes of soil- and woody peat categories, here usually characterized by higher ash content and often lower decomposition levels are additional contributing factors. The Moltakovsk peat deposit, almost exclusively based on a woody plant community as in the previous stage, continued to exist. Lakes continued to exist at various locations east of the peat deposit, among floodplain sediments along the rims of the earlier peat accumulation area. Small patches became overgrown by vegetation. Soil- and peat-formation took place here. Outside the ancestral Supsa River region the fluvial system preserved its previous configuration. Only the channel of the ancestral Khobi separated and acquired an independent outlet to the sea. Barrier ridges (facies MMV) formed in the central and northern parts of the plain. Their dimensions were comparable with those of the former barriers, as were the distribution characteristics of the underwater delta facies (MMD). North of the Nabadsk peat deposit, the retreating sea left lagoons behind.

As shown by absolute dates from the basal peat bed, during the 4.5-4 ka B.P. time interval (4530 ± 200- GIN-1416, Drillhole 380; 4480 ± 200- GIN-1133 m, Drillhole 382; 4800 ± 150- GIN-3621g, Drillhole 385; 4450 ± 150-GIN 1136, Drillhole 386), deposits of woody peat swamp Ispani-1 overlies sediments of the former sapropelic lake in the Colchis Lowland at all locations in the Kobuleti Basin. They formed an alder-bearing gelinitic-precollinitic peat, transitional along the margins toward a peat type with higher ash content and in a lower state of decomposition. Muddy zones (facies OZD) occurred in the swamp center [2], along with overgrown water bodies of forested floodplains (facies APZ) and soil formations (facies PDD). This indicates an unstable peat accumulating regime at the onset. The floodplain deposits north of the ancestral Kintrishi River occurred as far as the boundary of the sapropelic lake, completely covering the former mountain channel alluvium. At this time (4290 ± 55 [14]), the sea retreated from the area north of the mouth of the Natanebi River. According to Tvalchrelidze [16] a peat swamp formed here in the former marine shelf area and peat accumulation progressed eastward. Otherwise, the overall landscape pattern was inherited from the previous time interval, also including the underwater deltas in the ancestral Natanebi River region.

At 3 ka B.P., the landscape of the Colchis Lowland of the Poti Basin developed along the same lines as during the previous millenium. The sea in the southern part of the Poti Basin transgressed slightly further over the land area, covering a significant portion of the peat swamp in the Paliastomi lake-lagoon region. Sediments of the bay facies (MPZ) were deposited as the result. The extent of all the peat swamps (Churiya and Nabadsk swamps, the complex around Paliastomi lagoon) increased, as they extended their areas eastward. The distribution pattern of the earlier plant associations remained, except for *Sphagnum* concentrations that appeared in the central Imnatsk swamp area. To the north, the Anakliisk swamp obtained a dissected shape due to invasion of the lake by the forested floodplain (facies APZ), the overgrown lake facies (OZD) and the transitional category toward swamp facies (PDD). The area of the former Tikorsk swamp was covered by floodplain deposits (facies APV) of the ancestral Inguri River. Woody floodplain deposits formed (facies PDD) on its right-bank floodplain

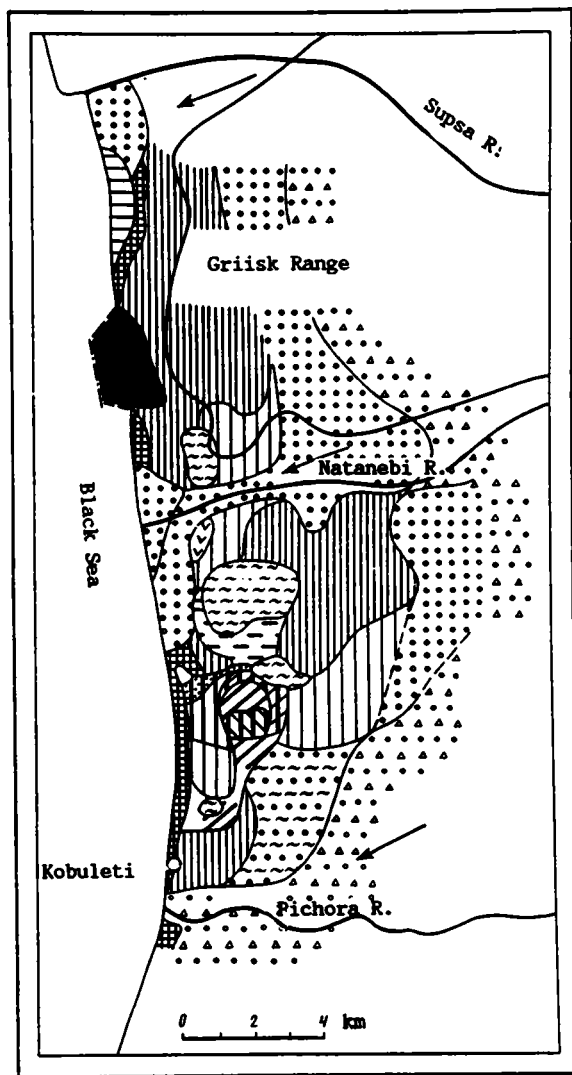


Fig. 5. Schematic paleogeographic map of Colchis Lowland, Kobuleti Basin, at c. 3 ka B.P. Symbols: Fig. 1.

area in the coastal zone, marking the beginnings of the Zоргatsk peat swamp. With exception of the ancestral Rioni River channel that had significantly shifted southward. Thus covering part of its former interior floodplain deposits, the drainage network in the area remained the same as before. Apparently, this was also related to subsidence of the Paliastomi region fault block. The ancestral Khobi River obtained its own channel network. The lagoon north of the Nabadsk deposit was preserved and the barrier ridges (facies MMV) became almost continuous. Locally, encroaching on the swamp, they increased in size. Small floodplain lakes were typical of the eastern area, occasionally overgrown by vegetation (facies APZ and OZD) and soil formations (facies PDD). They appeared and disappeared, and alternated during the following development stages, with deposits of paragenetically associated facies. The floodplain lake in contact with the Churiya peat swamp continued to increase in size as it kept growing eastward. The swamp's outlet was established as early as 6 ka B.P. An area of overgrown lake deposits (facies OZD) occupied a patchy area on the left-bank floodplain of the ancestral Rioni River. Facies that were transitional toward soil formations (PDD) were also present here. A large, locally overgrown lake existed on the Supsa River floodplain.

The Colchis Lowland landscape did not change in the Kobuleti Basin (Fig. 5) during this millenium. The ratio of mountain and lowland alluvium remained generally the same. The Is-pani-1 peat swamp slightly expanded beyond the boundaries of the previous sapropelic lake and locally included interior floodplain deposits. The previous woody plant communities in its central area became partially grassy and partially were enriched in *Sphagnum*. Woody peat formed along the swamp margin. These alternated with grassy components in the eastern swamp

area. A lake formed approximately in the center of the interior floodplain, north of the ancestral Kintrishi River. Its eastern floodplain bank zone has been overgrown by vegetation and later converted into peat swamp Ispani-2. The sea retreated even further on the north and was separated by a barrier ridge from the alluvium. The barrier in the area of peat swamps Ispani-1 and -2 widened and delta deposits of the ancestral Natanebi and Choloki Rivers increased in volume. A large lake developed on their floodplain (facies APZ), later overgrown by vegetation and converted into a swamp (facies PDP). Its southern area was represented by a facies combination of a woody peat swamp, with water circulation. It was located at the northern border of the Ispani-2 peat deposit. The peat swamp, located north of the Natanebi river mouth continued its existence.

At 2 ka B.P., the bay of the Black Sea became closed in the Paliastomi Lagoon area and the Pichora River swamp appeared in the Colchis Lowland of the Poti Basin. The Pichora River debouched into the developing lagoon. The Pichora peat swamp formed with its grassy and woody components, covering sediments of former overgrown lake (facies OZD) and related soil formations (facies (PDD)). The sea remained stable and displayed its former facies associations (MMM, MUM, MMD). Intensive sediment transport from the ancestral Rioni River accreted land areas north of the present Poti area and also contributed to the broad barrier ridge (facies MMV) that isolated the lagoon from the sea. The enrichment of the lagoonal aleurolites and clays in planktonic diatoms indicated that the lagoon's depth was greater at the same bay location than previously. The dominance of fresh water diatoms, despite the lagoon's brackish character, was due to the Pichora River runoff. Facies MMR represent fluvial deposits, delta deposits of small streams that entered the lagoon. Due to deposition of lagoonal deposits and associated facies types of the Pichora River, peat accumulation thus ceased in the entire Paliastomi lagoon area.

A map, not shown here, indicated the broader distribution of interfluvial floodplains of the ancestral Inguri-Churiya Rivers, due to significant reduction in their channel lengths. The ancestral Rioni channel was quite wide. Peat swamp dimensions increased somewhat in the same easterly direction that is most noticeable in the Imnatsk swamp area. Deposits of the same floodplain, including different types of lake bodies, distributed in this area, became swampy. Lakes developed on the Tikorsk swamp floodplain. As previously, the swamp centers were occupied by grassy peat, the marginal areas by woody types. A new peat swamp category of the Sphagnum association occurred in the Imnatsk swamp [2, 12], indicating an important stage in swamp development at the start of the Subatlantic period (2060 ± 150-GIN-1993; 2100 ± 150-Mo-251). The map shows the widespread distribution of barrier ridges (facies MMV), separating the swamp from the sea.

The main landscape change in the Colchis Lowland of the Kobuleti Basin at this time involved the widespread appearance of the sphagnum swamps in former grass swamp areas. As in the Poti Basin, these swamps were associated with the Subatlantic period, but on a larger scale. Woody plant associations were located along the rim zones of their former distribution area, forming woody, locally woody-grassy peats, often with high ash content. The other landscape units remained as before. Vegetative overgrowth of former lake bodies took place only north of the swamp complex and grassy swamp Ispani-2 became extended northward. The distribution areas of the Supsa, Natanebi and Choloki delta deposits were sharply reduced.

A thousand years ago the sea almost completely retreated from the Colchis Lowland in the Poti Basin. As in previous times, a "pseudoregression" influenced shoreline configuration. A marine sediment tongue remained only north of the Poti area. At this time, Paliastomi Lagoon became an open water lake, connected to the sea by the Kaparcha channel that cut across the coastal barrier (facies MMV). It overlaid earlier marine deposits and shallow marine, wave-influenced facies (MMM). This barrier represents a wide band that separated the Poti and Moltakvsk peat swamps and Paliastomi Lake from the sea, thus forming their western shore (facies MPP). The wide barrier formed at the area of the mouth of the ancestral Rioni River. The barrier sediments occurred in combination with floodplain facies (APP, APV) that divided the river channel into two distributary branches before reaching the sea. A subaerial delta formed. A small area was occupied by the Poti peat swamp and associated sediments of a floodplain that was converting into a swamp (facies PDD) between these distributaries. Portions of these facies formed even earlier. Thus, this time interval was characterized by widening of coastal barriers in the southern Colchis Lowland of the Poti Basin. In the northern half of the Lowland the barrier were more stable. Except for the following exceptions, the rest of the landscape features was almost identical to previously existing conditions (1) the appearance of the small Zoratsk peat swamp on the right bank of

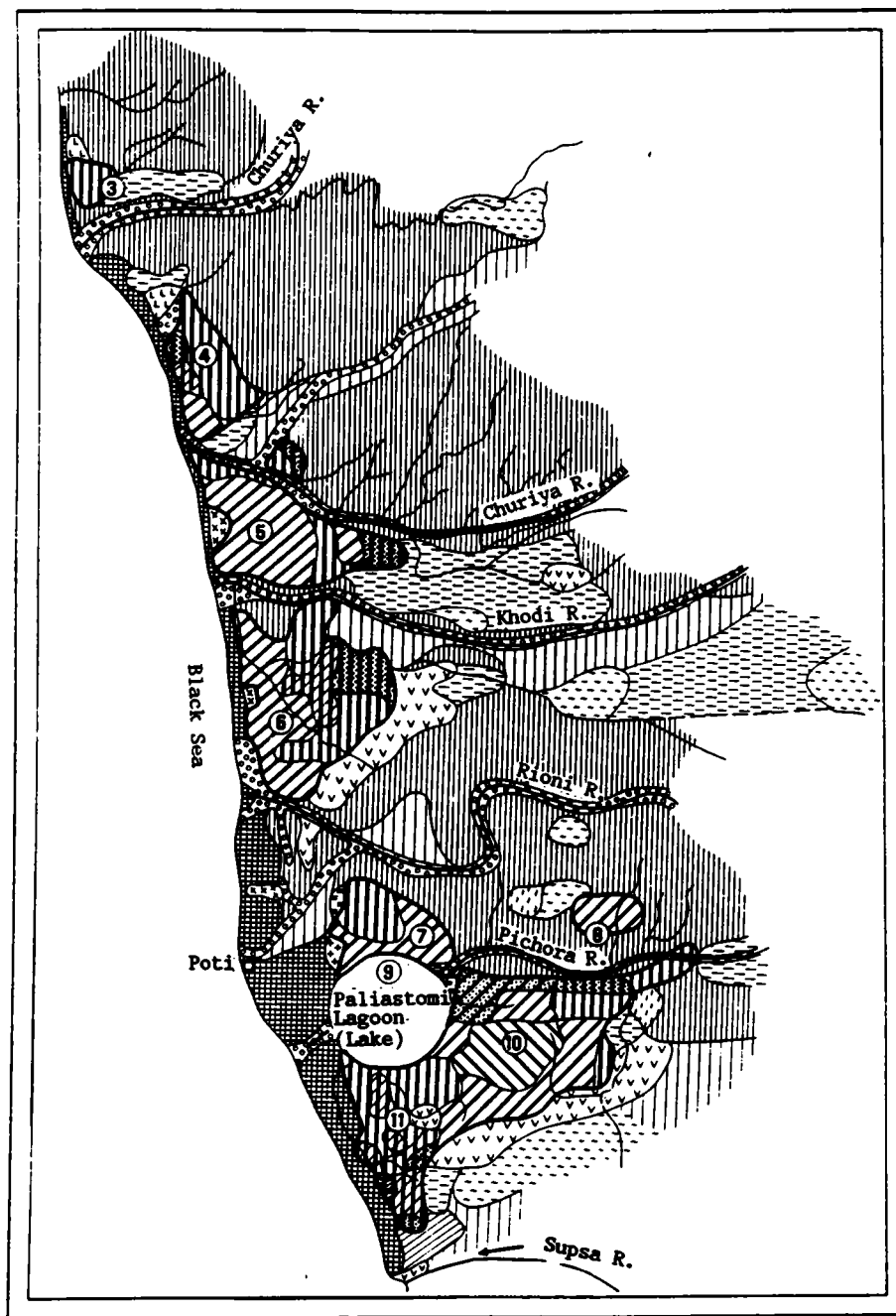


Fig. 6. Schematic landscape chart of the present-day peat accumulation zone, Colchis Lowland, Poti Basin. Symbols: Fig. 1.

of the ancestral Inguri River; (2) minor changes in the configuration of the Nabadsk peat swamp due to vegetative overgrowth and the conversion of floodplain deposits into swamp components, a process that progressively proceeded in a southeasterly direction; (3) slight widening of the Pichora and Imnatsk peat swamp area, due to vegetative overgrowth and conversion into swamp of the lower floodplain bank area of the Pichora River; (4) indications in certain areas that peat swamps were affected by siltation (facies OZD). This was especially noticeable in the southern half of the study area; (5) simultaneous decrease in the size of floodplain lake bodies (facies APZ), often a decrease in their size, as noted east of the Churiya and Tikorsk area swamps. The plant community distribution patterns in the peat swamps remained essentially the same but the sphagnum moss distribution area expands.

No landscape changes occurred in this time interval in the Colchis Lowland of the Kobuleti Basin. Only a large lake north of swamp Ispani-1, continued its northward growth and in the process of becoming a swamp.

The present landscape of the Poti Basin area (Fig. 6) inherited the previous overall landscape patterns while displaying changes in the configuration and dimensions of the landscape elements. The sea receded completely. Sediments of the coastal barrier ridge (facies MMV) extended as a separate unit, interrupted by rivers debouching into the sea. As before, it was widest in the south, especially around the town of Poti. The eastern marginal zones of peat swamp areas were reduced in size, and often effected by intensive siltation processes (facies OZD). This is clear in the SE Moltakvsk and Nabadsk peat swamp areas. Pichora Swamp, greatly reduced in size, disappeared completely in the Rioni Delta area. The plant community distribution patterns indicated that the previous condition continued.

Floodplain deposits in the Khobi and Churiya interfluvial zones and deposits of former lakes were converted into a soil-like state. The river configuration and their positions were inherited from the previous time interval.

The present Colchis Lowland in the Poti Basin landscapes (Fig. 7), in contrast with the previous configuration are characterized by the total absence of delta deposits in the Choloki, Natanebi and southern Supsa River areas, and marine deposits, covered by sediments that formed in coastal barrier ridges (facies MMV) and occupying a wide zone of a scalloped eastern outline. North of the Inguri River mouth, a peat bed is buried under barrier ridge deposits. As shown by the previous charts, a small peninsula jutted into the sea in the direct western continuation the peat unit. After the peninsula became covered by shelf waters, marine sediments of facies MMM started to form c. 3 ka B.P. In contrast with the previous landscape configuration, the barrier ridge in the southern Lowland increased its area in a northward direction, thus reaching the mouth of the Natanebi River. The relationship between the mountain- and lowland (including floodplain-) alluvium remained essentially stable. These deposits accumulated until 6 ka B.P. The northward growth of peat swamp Ispani-2 was typical. This came about through separation from the peat swamp Ispani-1 by means of a narrow floodplain strip (facies AAPV) that turned into a swamp in its western area (PDD). Sphagnum flora colonized the central swamp area, while the western and eastern margins were covered by woody-grassy, respectively, woody plant communities. The earlier plant community patterns and distribution characteristics remained in peat swamp Ispani-1. However, the sphagnum community expanded in all directions and covered the earlier formed woody and woody-grassy swamps. The large lake, that existed for a long time north of this swamp disappeared. Its deposits, mostly in the eastern areas, partially alternated with woody soils (PDD facies) and with floodplain deposits.

* * *

Comparative analysis of the described paleogeographic charts revealed patterns of change in the landscape units within various Holocene time intervals. They revealed evolution in the sediment accumulation patterns and also basic events during the paleogeographic evolution of the Colchis Lowland throughout the past ten millenia. Due to a "pseudoregression," during the Holocene the sea generally did not cover the Colchis Lowland. As shown by the charts, the areas of marine wave-effected (facies MMM), high energy shallow water deposits (bars, shoals, underwater ridges) and especially the delta facies (MMD), lenses of which totally disappeared in the Poti Basin c. 2-3 ka B.P., became steadily reduced in size. Deltas in the Kobuleti Basin disappeared somewhat later. The land actively prograded in areas of excess sediment transport and barrier ridges (facies MMV) formed. The barriers developed with great intensity c. 3 ka B.P. As shown by chart comparison, the described basic changes in the patterns of marine facies complexes with time and in their distribution patterns were accompanied at the same time by changes in the relationship between the continental landscape units. The floodplain alluvium, uniform in Old Black Sea times in the Colchis Lowland of the Poti Basin, by New Black Sea times gradually became differentiated into channel- and floodplain alluvium categories. The same happened with the New Black Sea mountain alluvium of the Colchis Lowland in the Kobuleti Basin. At the time of this gradual differentiation, the mountain alluvium became transitional toward floodplain alluvium in the peat accumulation area. It was preserved only in its marginal zone and the southernmost parts in the Kintrishi River region. As the charts show, the differentiation of alluvial zones was accompanied by the narrowing of migrating channel zones. The areas of the interfluvial belts, including the near-channel floodplain zone and the internal floodplains increased, as the result. This is how the ancestral Rioni River and its right-hand tributary, the ancestral Khobi River first

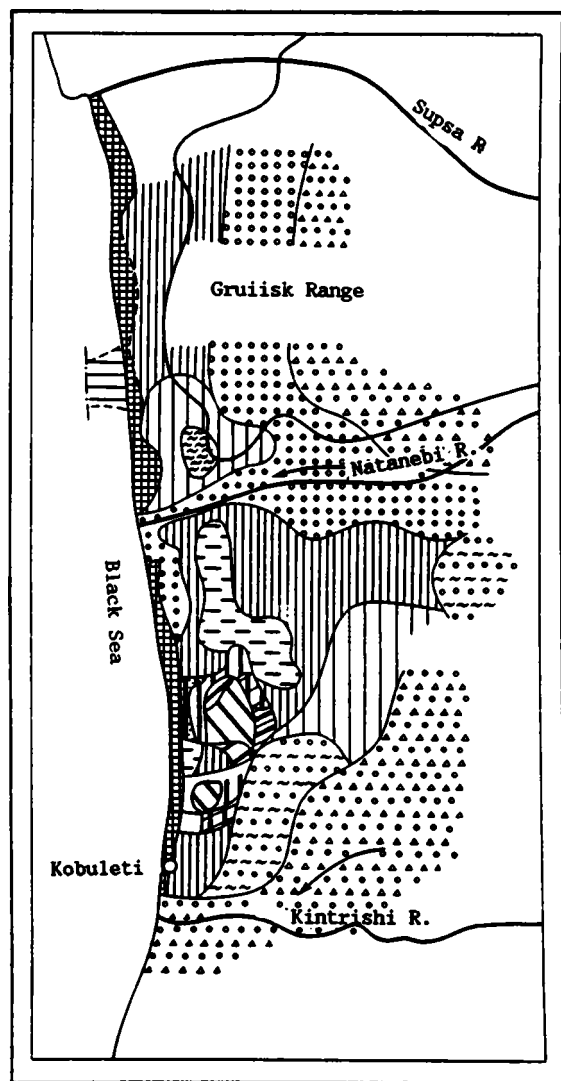


Fig. 7. Schematic landscape chart of the Colchis Lowland at present. Symbols: Fig. 1.

appeared in the landscape c.6 ka B.P. The channel of the ancestral Supsa River evolved and terminated in the Black Sea SW of the present Paliastomi Lagoon. Landscape elements of that river's floodplain also developed at the time. Channels of the ancestral Natanebi and Kintrishi Rivers also became manifested. Finally, the main rivers of the Colchis Lowland fluvial network occupied approximately their present positions only about 2-3 ka B.P. The channel of the ancestral Rioni River shifted southward c. 3 ka B.P. and at the same time the ancestral Khobi River became a separate stream, starting 4 ka B.P., the channel of the ancestral Supsa River shifted significantly (7-8 km) southward. About 2 ka B.P. the swamp stream Pichora appeared, emptying its waters into the Paliastomi River. The change in stream positions in the southern half of the Colchis Lowland in the Poti Basin coincides with the maximum extent of the Black Sea transgression 2-4 ka B.P. It is clearly defined through analysis of the charts that indicate positions of bay (MPZ) and lagoon (MPL) deposits, covered by Supsa River floodplain deposits and peat in the Paliastomi Lagoon area. West of this location the sediments are overlain by facies MMM; high wave energy marine basin deposits.

The episode of interfluvial floodplain area broadening c. 6 ka B.P. was associated with appearance of small areas of swampy landscape, lakes (facies APZ), occasionally vegetation-covered (facies OZD, OZT), and sapropelic lakes (facies OSO, OSS, OSP; Kobuleti Basin), indicating the start of peat accumulation in the Colchis Lowland. Peat commenced to form first in the areas of the Churiya, Nabadsk, Paliastomi and Imnatsk peat deposits; in the lowest parts of the Colchis Lowland in the central Poti Basin area. Charts show that continued development was associated with the widening of peat swamp areas in different direction to c.

3 ka B.P. and the appearance of new swamps, as the Anakliisk Swamp five millenium ago. Later, the Tikorsk Swamp formed along the northern rim of the Poti Basin, and in 4 ka B.P., the Gagidansk and Abkhaz-Mergel'sk Basin deposits. Peaty swamp "Ispani-1" in the Kobuleti Basin formed c. 4-4.5 ka B.P. Somewhat later, "Ispani-2" formed. At the start of the third millenium, the Poti Basin peat swamp zone widened only eastward, including floodplain and lake deposits. The area of these facies included, also in their northern and southern extension, river channel deposits. Westward, a quite wide coastal barrier ridge formed (facies MMV). The widest peat swamps that developed 1-2 ka B.P., form an almost continuous blanket along the coastal zone of the Colchis Lowland, interrupted only by streams that empty into the sea. At the same time, new peat swamps also appeared; such as the Pichora Swamp on the right bank of the Pichora River and the Zogratsk Swamp south of the Abkhaz-Mergel'sk Basin at the mouth of the Inguri River. Starting c. 1 ka B.P. and continuing through the present swamps became reduced in the size in the Poti Basin due to mud accumulation mostly in the SE marginal zone. In contrast with the Poti Basin, the peat swamps did not significantly change in size, only swamp "Ispani-2" expanded slightly northward.

Analysis of the charts indicates the presence of plant communities and their distribution patterns during the evolution of the swamp landscape. The start of peat formation in the Colchis Lowland everywhere is related to woody plant communities that later dominated only along the rim of peat swamps and in the marginal zone of the Poti Basin [2]. Grassy plant communities developed without changes in the central portion of the peat swamps down to the coastal zone. Only in small portions of Imnatsk Swamp of the Poti Basin are they transitional toward a mossy community, present throughout modern times. They occurred almost everywhere in the Kobuleti Basin (Ispani Swamps 1 and 2) since the start of the Atlantic climatic period (almost 2 ka B.P.).

Analysis of the chart also shows that during its entire history, as far westward as the shoreline, marine ingression did not interrupt peat accumulation in the central peat swamp area. The Paliastomi Lagoon area is an exception; traces of bay-lagoonal deposits were left here behind during the peak of the New Black Sea transgression (facies MPZ and MPL). As shown earlier, they covered the peat bed c.4 ka B.P. Such a stable regime of peat accumulation indicates that no significant Black Sea transgressions could have taken place in the Poti Basin peat deposition area after 4 ka B.P. Such transgressions were mentioned in the literature [4, 16, 17] with regard to this area and other Black Sea Basin areas.

It should also be noted that the present Paliastomi lake-lagoon is a relict of the Black Sea from the New Black Sea times. Its formation started c.4 ka B.P. in a swampy region and was completed c.2 ka B.P. This fact conflicts with earlier assumptions [11, 12] that considered the lagoon remnant of Colchis Bay of the Black Sea. The Bay did not exist prior to the Old Black Sea stage. It was not present prior to peat formation that is now dated as having been started prior to 6 ka B.P.

Review of the paleogeographic charts, covering the time period between the Old Black Sea stage and the present, the alteration pattern in the interrelationship of landscape units in time and space, was based on the distribution- and configuration characteristics of these units, as well as the alternation between different forms that are not paragenetically inter-related. This pattern became the tool for the prediction of what landscape units will emerge in the nearest future in the Colchis Lowland. It may already be predicted that while retaining the present general geological setting in the Colchis Lowland, the area of peat accumulation and main peat swamp landscape units gradually will diminish in size. At the same time, based on the principle of uniformitarianism [6], the landscape unit interrelationship pattern and the paleogeographic charts of Holocene Colchis Lowland peat swamp deposits provided conclusive evidence for the accuracy of assumptions on paleolandscapes and the genetic conditions of ancient deposits, under which corresponding types of coal-bearing formations existed in intermontane basins, open to the sea.

LITERATURE CITED

1. I. P. Balabanov, A. M. Kvirkveliya and A. B. Ostrovskii, Latest History of the Formation of Engineering-Geological Conditions and Long-Term Prediction on the Evolution of the Coastal Zone of the Pitsunda Peninsula [in Russian], Metsniereba, Tbilisi (1981).
2. L. I. Bogolyubova and V. A. Kotov, "Sediment genesis in the peat accumulation area of the Colchis Lowland in the Holocene," *Litol. Polezn. Iskop.*, No. 5, 37-38 (1989).
3. Ch. P. Dzhanlidze, "Paleogeography of the Colchis Lowland in the Late Pleistocene," *Izv. Akad. Nauk SSSR, Ser. Geogr.*, No. 1, 63-65 (1975).

4. Ch. P. Dzhanelidze, Holocene Paleogeography of Georgia [in Russian], Metsniereba, Tbilisi, (1980).
5. Yu. A. Zhemchuzhnikov, and V. S. Yablokov, L. I. Bogolyubov, L. N. Botvinkina, Composition and Accumulation Conditions of the Middle Carboniferous Basic Coal-bearing Formations and Coal Beds in the Donets Basin [in Russian], Part 1, Izd.-vo Akad. Nauk SSSR (1959).
6. Yu. A. Zhemchuzhnikov, "The question and present state of the uniformitarian method in lithology," Litol. Sb., No. 1, 59-66, Gostoptekhizdat, Leningrad (1948).
7. Yu. A. Zhemchuzhnikov, "What is facies?" Litol. Sb., No. 1, 50-58, Gostoptekhizdat, Leningrad (1948).
8. E. V. Kvavadze, "New data on the stratigraphy and paleogeography of the Colchis Lowland Holocene," in: The Quaternary System of Georgia [in Russian], Metsniereba, Tbilisi, 123-129 (1982).
9. A. G. Laliev, "The geology and development history of the Colchis Lowland," Trudy GIN Akad. Nauk Georgian SSR, Ser. Geol., Vol. 10 (15) (1957).
10. A. V. Makedonov and T. A. Ishina, Basic Principles of the Composition and Formation of Coal-bearing Formations and Methods of Predicting Coal Content [in Russian], Nedra, Leningrad (1986).
11. A. V. Motsereliya, "Geological history of the Colchis Lowland," Bull. All-Union Sci. Inst. Tea Subtropical Cultivation, No. 3, 110-115 (1950).
12. M. I. Neishtadt, "Study results of Holocene deposits," in: Paleogeography and Chronology of the Upper Pleistocene and the Holocene [in Russian], 112-132, Nauka, Moscow (1965).
13. M. G. Tvalchrelidze, "New data on the paleogeography of the southern Colchis Lowland," in: Scientific Symposium in Honor of the 70th Anniversary of the October Revolution, 76-77, Metsniereba, Tbilisi (1987).
14. M. G. Tvalchrelidze, Lithology of Recent Sediments of the Southeastern Black Sea Coastal Zone [in Russian], Candidate Dissertation, Geol.-Mineral. Sciences, GIN, Akad. Nauk Azerbaidzh. SSR, Baku (1987).
15. P. P. Timofeev, Jurassic Coal-bearing Formations of Southern Siberia and Their Formation Conditions [in Russian], No. 198, Nauka, Moscow (1977).
16. P. V. Fedorov, "Late Quaternary history of the Black Sea and developments of the southern seas of Europe," in: Paleogeography and Pleistocene Deposits of the Southern Seas of the USSR [in Russian], 25-31, Nauka, Moscow (1977).
17. D. V. Tsereteli, Pleistocene Deposits of Georgia [in Russian], Metsniereba, Tbilisi (1966).
18. D. V. Tsereteli and N. S. Mamatsashvili: "New data on Middle and Upper Pleistocene deposits of the Colchis Lowland, Black Sea coast," Bull. Commission for Quaternary Period Studies, 26-37, Nauka, Moscow (1975).

BASIC PATTERNS OF TERRIGENOUS SEDIMENTATION AT THE MOUTHS OF RIVERS IN ARID AND SEMIARID ZONES

Yu. P. Kurustalev

UDC 552.5:551.585.53

Studies have shown that very powerful mechanical differentiation takes place at the mouths of rivers in arid and semiarid zones (Volga, Don, Kuban'), as a result of which fine-grained particles are separated from coarse. In the fore-delta area, where the zone of mixing of river waters with marine generally takes place, sudden decrease in the current velocity, along with an extremely high rate of deposition, result in the accumulation of poorly sorted terrigenous bottom sediments that are similar in composition to alluvium, with a very wide range of grain sizes, from sand to fine silt and even clay.

The delta regions of rivers are characterized by a unique depositional environment. They are very complex natural systems where there is a mixing of the hydrologic, hydrochemical, and hydrobiological conditions of sedimentation of rivers and the sea. This is an arena of intensive reorganization of sediments derived from the continent; of the interaction of fresh and marine waters that vary in chemical composition; of complex biogeochemical processes; and of massive deposition of elastic terrigenous material. Here, in a position intermediate between river and marine systems, sedimentary processes have aspects of both freshwater and marine deposition. The accumulation of the majority of the terrigenous material at mouths of rivers was classified by Lisitsyn [7] as due to submarine slides. All of this has an effect on the way sedimentary deposits are formed, determining their complexity and variety, and shows river-mouth areas to be unique subjects for elucidation of patterns of modern sedimentation, without which it would be impossible to understand marine sedimentation as a whole.

It has been established in recent years that such areas are depositional basins for gas and oil deposits. Their origin is of special interest to the petroleum geologist. Paradoxically, however, sedimentational processes (especially hydrochemical and biogeochemical) in deltas and along seashores at river mouths have been poorly studied. As a result of this situation, there is a need for description of these features and the basic patterns of deposition at the mouths of rivers entering epicontinental seas in arid and semiarid zones: Azov, Caspian, and Aral'.

Among the aspects of sedimentation in arid epicontinental seas, of particular note is the considerable area of river mouth regions. The total area of the deltas and shores of the Don and Kuban amount to more than 14% of the Sea of Azov, and if one includes the Gulf of Taganrog, actually an estuary of the Don, the figure increases to 27.5%. Similarly, the area of the mouth of the Amudar'ya River alone amounts to 21.1% of the sea of Azov, and a considerable part of the North Caspian Sea is the delta region of the Volga, Ural, and Terek Rivers (Table 1).

SUPPLY OF FLUVIAL SEDIMENTS

Epicontinental seas in arid zones are characterized by many different sources of supply and an abundant influx of sediments [23]. The principal suppliers of these sediments are rivers (Table 2). Data is included for the Sea of Azov, where as a result of an unusual geological setting and the composition of the rocks, the products of erosion are particularly important.

The watersheds of the rivers entering the seas under study can be subdivided geographically into two zones: development of the load and its removal. Where precipitation is considerable in the first zone, there is extensive weathering and erosion of the bedrock. This

Rostov State University. Translated from *Litologiya i Poleznye Iskopaemye*, No. 1, pp. 91-104, January-February, 1990. Original article submitted August 9, 1988.

TABLE 1. River Mouth Areas in the Azov, Caspian, and Aral' Seas

River	Areas, 10 ³ km ²			Percent of total area of Sea
	delta	shoreface	river mouth area	
Don	0,54	0,50	1,04	2,8
Kuban	4,30	0,10	4,40	11,9
Ural	0,60	0,10	0,70	<u>0,2</u> 0,8
Volga	18,95	5,34	24,29	<u>6,5</u> 27,9
Terek	8,88	3,69	12,57	<u>3,4</u> 14,4
Kura	0,10	0,65	0,75	0,2
Amudar'ya	8,99	4,95	13,94	21,2

Note: numerator is for area in entire Caspian, denominator is Northern Caspian.

TABLE 2. Sediment Supply to Epicontinental Seas in Arid Zones

Principal source	Sea of Azov		Northern Caspian		Aral' Sea	
	mill. tons	%	mill. tons	%	mill. tons	%
Solid load*	7,70	14,7	32,15	24,4	82,30	58,6
Dissolved load	11,47	22,0	74,00	56,1	22,21	15,8
Eolian material	5,00	9,6	25,67	19,5	36,00	25,6
Material eroded from shore and sea floor	28,00	53,7	—	—	—	—
Total	52,17	100,0	131,82	100,0	140,51	100,0

*Data is for natural flow.

is where the sediments are formed that end up in the basin of deposition. In regions with an arid climate, however, the runoff is decreased. Here the river gradient is generally quite small, the current is slow, and sand bars are common. The relatively low capacity for transport results in the accumulation of a considerable part of the terrigenous material, mainly coarse-grained sediment, in the zone of load removal.

The ratio between the area of development of the load and the area of its removal varies. The highest values are for the Volga, Don, Kura, Ural, and Terek rivers (5:1 or more), and the lowest for the Amudar'ya (1:1) and Sydar'ya (1:2). The watershed of the Amudar'ya covers 227,800 km² and is located in the Pamir-Alai Mountains, a highly dissected terrain with elevations up to 6000-7000 m or more [30]. The sediment load in the Syrdar'ya basin is formed in the mountains and foothills of the Tian-Shan system and in part the Pamir-Alai system, surrounded by the Fergana lowlands and the Golodnii and Dal'verziisk steppes. The mountainous part of the watershed covers 150,000 km². The area of the river with elevations of less than 1000 m, located on the Hercynian platform, almost never contributed to the discharge. The area of load removal of the Amudar'ya covers 237.2 thousand km², and of the Syrdar'ya, 310,000 km² (significantly greater than the area of load formation). This ratio determines the amount of river water and sediment lost during transportation. Using data on turbidity collected over many years, Rogov, Khodkin, and Revina [14] obtained an average suspended load on the Chatly section (before control of the Amudar'ya) of 129 million tons within a range of 69 to 183 million tons. A huge mass of this sediment (about 60.2 million tons) was deposited on the delta [30]. In the area from Volgograd to the Caspian Sea, which is located in an arid climate, the Volga has lost as much as 50% of its suspended load [4]. Analysis of the distribution of the discharge of sediments in the Kuban shows that about 48% is deposited in the lower reaches, 25.6% between Krasnodar and the delta, and 22.4% on the delta. Broad deltas, large areas of swamps and distributaries, and silting of channels restrict the

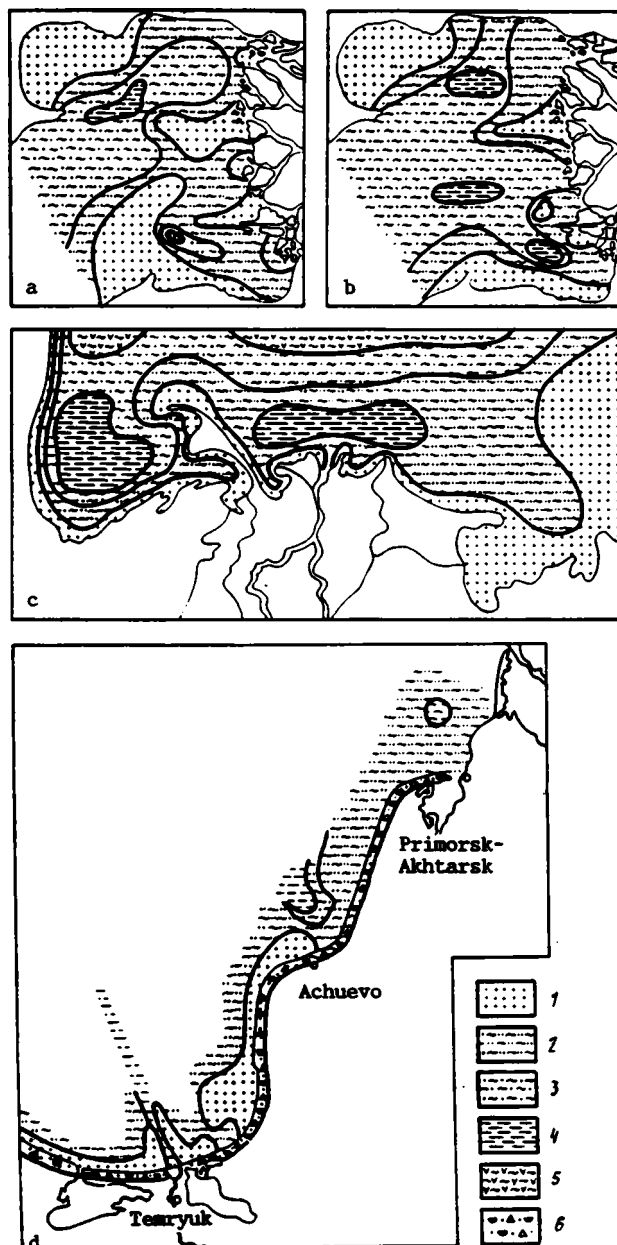


Fig. 1. Distribution of lithotypes in bottom sediments along the shoreface at the mouth of the Don after offshore (a) and onshore (b) movement of water (after [3]), and at the mouth of the Amurda'yia (c) and Kuban' (d): 1) sand; 2) coarse silt; 3) fine silty mud; 4) clayey mud; 5) clayey-limy mud; 6) shells and detritus.

sediment load of other rivers as well. The Syrdar'yia and Ural rivers deposits about one third of their suspended load in their estuaries, and the Don, as much as 40% (12% during natural flow).

The fine grain size of the suspended sediments in the rivers under study is the result of the particular geologic makeup, the lithologic composition of the rocks in the watershed area, the presence of bedrock in the transport zone, and the small amount of forested areas. In the grain size classification of Shamov [27], the lower Kuban', Ural, and Emba may be referred to the first category (90% of the grains less than 0.05 mm), while the Volga, Amurda'yia, Syrdar'yia, Don, and Terek are in the second category (75-90%). The lower reaches of the smaller rivers, whose watersheds lie entirely within the Caucasus Mountains and the

TABLE 3. Change in the Discharge and Solid Sediment Load at Volgograd after Control of the River Flow

Year	Liquid discharge		Solid load	
	km ³	%	mill. tons	%
Before control of flow				
High water years (1946, 1947 1953, 1955)	272,5	100	19,3	100
Av. years (1950)	227,0	100	11,0	100
Low water years (1949, 1951, 1952, 1954)	218,0	100	10,7	100
After control of flow				
High water years (1957, 1958)	265,0	97	16,0	83
Av. years (1961, 1962)	232,5	102	9,1	83
Low water years (1955, 1959 1960)	214,0	98	7,3	68

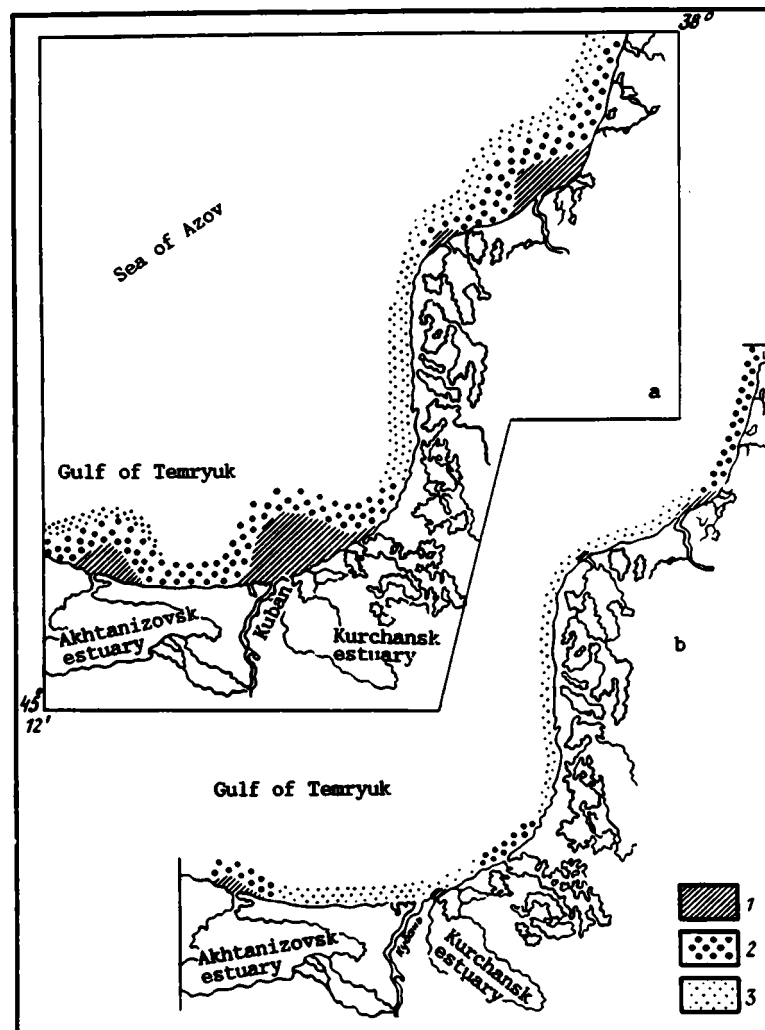


Fig. 2. Distribution of suspended sediments from the Kuban resulting from southeasterly (a) and westerly to southwesterly (b) winds [29]: 1) brownish-yellow Kuban water; 2) intermixed river and marine waters; 3) greenish-yellow marine water.

TABLE 4. Rate of River Current and Content and Grain Size Distribution of Suspended Sediment during Calm Weather with Distance from the Mouth of the Don [3]

Distance from mouth, km	Rate of current, cm/sec	Suspended sediment, mg/l	Size range of suspended sediment (numerator, mg/l; denominator, %)				
			> 0,1	0,1-0,05	0,05-0,01	0,01-0,005	< 0,005
Mouth	72	124	5,7 4,6	26,4 21,3	50,0 40,3	4,2 3,4	37,7 30,4
1	44	109	2,1 1,9	13,3 12,2	39,9 36,6	13,8 12,7	39,9 36,6
2	15	86	0,3 0,4	2,0 2,3	10,7 12,4	23,8 27,7	49,2 57,2
3	8	47	0,1 0,2	0,3 0,6	3,1 6,7	12,1 25,7	31,4 66,8
4	5	17	- 0,1	0,1 0,3	1,0 6,0	4,0 23,8	11,9 69,8
5	4	14	- -	0,1 0,2	0,7 5,3	3,1 22,4	10,1 72,1

Azov crystalline massifs, characteristically have a high sand and coarse silt content (greater than 50%) and belong to the third category.

In the last two decades the watersheds of rivers in the southern USSR have been drastically affected by man. As a result of construction, reservoirs, and other hydrotechnical developments, there have been substantial changes in the hydrologic regimes, the volume of solid sediment, the grain size distribution of suspended sediment, and the chemical content of the river water, i.e., the principal factors that determine conditions of deposition and the composition of the sediment. Forty years ago the solid load at the upper end of the Volga delta amounted to 25.7 million tons [8]. After control of the flow, the suspended material amounted to a total of 9.1 million tons in a year of average discharge (Table 3), the bed load was reduced to 0.8 to 2.5 million tons, and the average particle size decreased (0.4-0.5 mm in 1934 and 0.14-0.17 mm in 1958).

Even greater effects have been observed in the Kuban basin, associated with the construction of the Krasnodarsk reservoir. The suspended load has decreased ten times in comparison with the natural flow (from 7.54 to 0.83 million tons).

According to Simonov [15], at the last sections of the estuary the total sand and siltstone fraction previously amounted to 22-27%, which resulted in the contribution to the foredelta area of about 1.2 million tons of sediment per year. After control of the river flow, the concentration of coarse particles in suspension decreased as much as 10-15%, and currently only 0.1 to 0.2 million tons are being deposited on the shoreface off the Kuban. Construction of the Tsimlyansk hydrologic system on the Don River led to almost a twofold decrease in the terrigenous sediment load. After completion of the Mingechaursk reservoir, the solid load of the Kura decreased 70% [28].

PATTERNS OF TERRIGENOUS SEDIMENTATION

The interaction of river flow with waters of the basin is the main process of mechanical differentiation at the river mouth. This interaction results in the dispersal of the river energy and the consequent loss of the capacity to transport alluvium [6, 11]. The relationship between the intensity of sedimentation at the shoreface of the Don and the transporting capacity of the current is shown in Table 4. During high water in calm weather, most of the sand and coarse silt drops out of suspension onto the surface of the foredelta, while the fine silt accumulates on the bar and on the seaward slope, along with some of the coarse pelitic material [3]. Because of their small hydraulic size, fine and medium silt fractions pass through the upper part of the shoreface and are deposited mainly at its outer margin (see Table 4). As a rule, the transport capacity of the river current on sedimenta-

TABLE 5. Rate of Deposition of Terrigenous Material in the Volga Delta and Shoreface

Study area	Area, km ²	Av. rate of deposition, mm/yr	Volumetric weight, g/cm ³	Moisture, %	Rate of deposition of terrigenous material		Terrigenous material carried beyond the delta, mill. tons
					mill. tons/yr	recalc. as carbonate-free	
Delta	18,9	0,63	1,68	27,8	14,2	12,7/44,9	15,6
Delta from							
depths to 2 m	12,9	0,63	1,57	31,2	8,8	7,2/25,4	8,4
depths to 3 m	19,7	0,63	1,57	31,2	13,4	11,0/38,9	4,6

Note: Rate of deposition since NeoCaspian time determined by data from more than 300 holes drilled in the Volga delta and shoreface (average thickness of NeoCaspian sequence, determined by radiocarbon dating to have been deposited over the last 6000 years, is 3.7 m on the delta and 3.8 m on the shoreface). 2. Volumetric weight and moisture content determined from one hundred analyses by Soyuzmorniiproekt. 3. Average carbonate content in delta, 10.8%; in shoreface, 18.1%. 4. Suspended material introduced into delta area during natural flow, 25.0 mill. tons; bed load, 3.3 mill. tons. 5. Numerator, mill. tons, and denominator, percent of solid load of the Volga.

TABLE 6. Size Distribution of Don Sediments and Shoreface Deposits, %

Sample location	Size fraction, mm			
	> 0,1	0,1-0,05	0,05-0,01	< 0,01
Razdorskii Station	14,0	10,1	30,7	44,2
Center of river mouth (town of Azov)	4,0	7,3	27,8	60,0
Shoreface deposits (av. content)	21,3	10,8	37,6	30,3

tion only affects the foredelta. For example, the current velocity of the Volga decreases ten times from the mouth to deep water (from 1.3-1.5 to 0.15 m/sec), and of the Syrdar'ya, during natural flow, 15-20 times (from 1.6-2.0 to 0.1-0.15 m/sec) [18]. The role of the river water is to force out marine water and preserve its hydrodynamic and chemical properties, which are gradually lost toward the open sea [20]. The boundary of active deposition for the Bakhtemirsk branch of the Volga is 130-140 km from its mouth, where the depth is generally 6-7 m. The flow of the Terek is felt out to 10 km, at a depth of 10-12 m, and the Don out to 4-5 km.

The sediment that accumulates on the bottom of epicontinental seas may be further affected by wind and waves. As shown by Bronfman and Aleksandrov [3], when winds drive the water offshore the velocity of bottom currents of the Don increases, which causes erosion of submarine ridges and deposition of sand and coarse silt on the slope; fine silt and pelitic sediments are removed from the foredelta so that it is left with coarser grained material (Fig. 1a). Onshore winds promote the accumulation of sediments with little sorting on the shoreface and on areas off concave, protected parts of the delta margin (see Fig. 1b). Thus on the Don foredelta, where river and marine waters generally intermingle, intermittent decreases in the current velocity along with extremely rapid deposition result in the accumulation of a broad range of sediments, from coarse sand to clayey mud.

The wave regime is a principal factor in sedimentation off the mouth of the Kuban. The quantity and distribution of suspended sediments introduced by the river depend on the intensity and trend of the waves. Generally south and southeast winds are most favorable for

TABLE 7. Variations in Solid Load and Deposition for Different size Fractions on the Delta and Shoreface of the Don, Mill. Tons (during natural flow)

Fraction, mm	Solid load		Deposition		Carried beyond the shoreface
	Razdorskii	Azov*	on delta	on shoreface †	
> 0,1	0,79	0,34	0,45	0,38	0,07
0,1-0,05	0,40	0,25	0,15	0,19	0,06
0,05-0,01	1,21	0,97	0,24	0,67	0,30
< 0,01	1,74	2,08	-0,34	0,54	1,54
Total	4,14	3,64	0,50	1,78	1,97

*Data from Mouth-of-Don gaging station.

† Deposition on shoreface (500 km²) determined from sedimentation rate, using data from Panov and Mamykina [12].

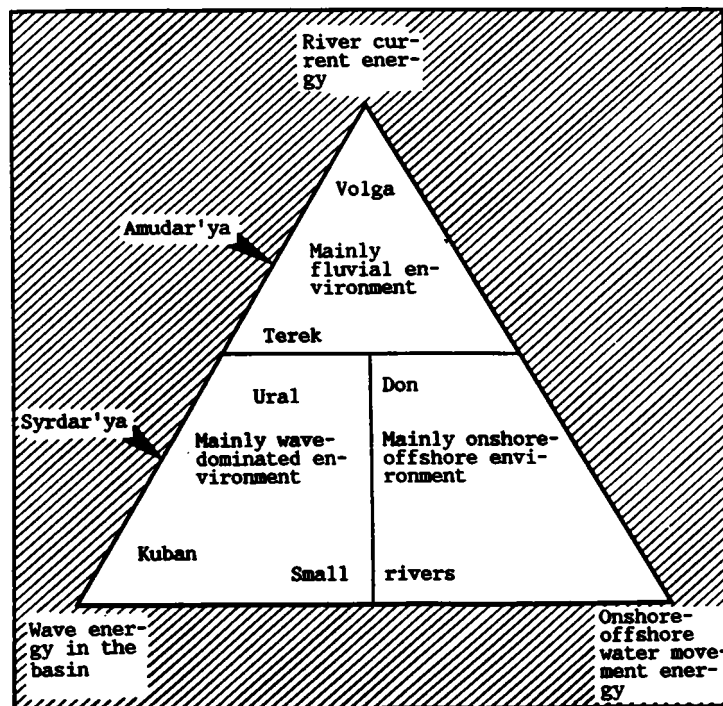


Fig. 3. Classification of river mouths in nontidal seas according to hydrodynamic energy, which determines the types of sedimentation.

movement of freshwater into the Sea of Azov (Fig. 2a). During these winds, high concentrations of suspended sediment from the Kuban are located 10-12 km out in deep water [29]. A different situation arises when the winds are from the west. In this case, suspended sediments are carried no more than 1 km from the delta margin. Only a narrow strip along the shore is affected by the solid and liquid discharges (Fig. 2b).

On the foredelta, and especially beyond it, deposition of pelitic material fosters coagulation and, on shorefaces in arid climates, the formation of clayey-carbonate clots. According to Skopintsev [17], the most favorable conditions for such aggregations is with high concentrations (250-300 mg/liter) of suspended sediment and a critical salinity of about 2‰. This is supported by the sharp decrease in the amount of pelitic clasts in suspension in the river mouth area. In particular, on the shortface of the Don clay particles decrease by a factor of two within a small distance (less than 1 km). Apparently, in the more mineralized marine waters the negative charge on the surface of clay minerals is neutralized by adsorption of chemical elements, decreasing forces of repulsion, which promotes aggregate formation [32, 33]. The destabilization of fluvial suspensions is active at



Fig. 4. Types of deltas: multi-distributary (Volga, Terek, Amudar'ya, Don); filled (Kuban); single-channel (Syrdar'ya, Kura).

salinities up to 6%. Further increase in salinity sharply decreases the coagulation effect. Flood waters from coagulation aggregates that average 0.0035 mm in size [17].

The foregoing information supports the conclusion that the principle form of clay minerals during transportation in the marine environment is as aggregates and not as discrete particles [21, 24]. The capability of coagulating the forming aggregates large enough to be affected by gravity, and this mechanical sorting, is not the same for different minerals. Clay minerals can be ranked as to their ability to aggregate and their hydrodynamic activity in the area of rivers mouths in the following order: kaolinite-chlorite-mixed layer clays-montmorillonite [22]. Sorting effects in the marine environment have been demonstrated experimentally. Dyer [31] has thus shown that kaolinite and illite are most completely deposited in salinities as low as 4%, while montmorillonite coagulates in a wider salinity range.

The formation of aggregates is typical of river-mouth regions in humid as well as arid zones. Deposition of large masses of fine-grained material takes place in the Baltic Sea directly adjacent to river mouths as a result of the formation of aggregates in hydraulic sizes of such minerals as montmorillonite and mixed layer clays. A considerable quantity of kaolinite is deposited in the upper reaches of Pamlico Sound, while illite is stable with respect to coagulation and passes through the sound [32].

Suspended sediment in arid zone deltas includes, in addition to clay mineral aggregates, clayey-carbonate clots that contain authigenic and terrigenous calcium carbonate. The coagulation of pelitic suspensions and the formation of chemogenic calcite are essentially areally coextensive. The later forms most readily, according to microscopic data, in salinities of

TABLE 8. Discharge of Water, km³ (in numerator) and Solid Load, Mill. Tons (in denominator) from the Kuban (compiled from data from the mouth of Kuban gauging station)

Year	Krasnodar	Head of delta (Tikhovskii)	Mouth of Kuban (Zaitsevo bend)	Mouth of Pro-toka (Slobodka)
1965	12,4/9,46	11,3/6,62	5,9/2,81	5,0/4,10
1966	12,4/8,20	11,1/6,31	5,6/2,52	3,9/2,71
1967	13,2/11,67	12,4/7,57	5,4/3,09	4,3/2,90
1968	15,0/9,17	14,1/6,32	6,6/2,31	4,9/1,58
1969	6,8/3,78	5,9/2,24	3,0/0,82	1,9/0,29
1970	11,0/5,68	10,5/5,05	5,4/1,61	4,3/1,10
1971	9,9/4,42	9,0/3,47	4,6/1,07	3,2/0,73
1972	12,4/7,91	11,0/7,27	5,5/1,99	4,2/1,68
Av.	11,6/7,54	10,7/5,61	5,3/2,03	4,0/1,89

1.5 to 5%. The formation of such clots is promoted to a considerable degree by adsorption of calcium ions by kaolinite and hydromica and by the degradation of clay minerals as a result of changes in the sedimentary environment. Coagulation accounts for 20-30% of the total volume of terrigenous material of the Kura, and 10-20% for the Kuban [5, 16]. Our calculations have shown that, during calm weather, up to 73% of the pelitic particles in the Don estuary have been deposited (see Table 4).

It is also important to note that fine silt and clay deposits are not formed throughout the region of the delta, but as a rule, only along the main outflow of the river. This is true of the Volga, Amudar'ya, Don, and Ural, which are characterized by gentle bottom relief and large flows of water in the main channel or larger distributaries (see Fig. 1a-c). Where the delta has a steep slope into deep water and numerous distributaries, the areal impact of the solid load on sedimentation is limited, and fine grained sediments are found in isolated areas or are essentially lacking. This type of distribution of bottom sediment lithologies can be seen in the shoreface of the Kuban (see Fig. 1d). An imprint on the sedimentary mechanism in delta regions is left by not only loss of energy, decrease in velocity of the river current, and aggregate formation, but also by the grain size distribution of the alluvium, the morphology of the submarine slope, the shape of the shoreline, and other factors.

The studies carried out have made it possible to estimate quantitatively the rate of deposition of terrigenous material in the river-mouth regions of the Volga, Don, and Kuban. In the Volga delta (one of the most highly branched and morphologically complex), with more than 500 distributaries and passes, more than 44% of the clastic material delivered by the river has been deposited since Neocaspian time. Depositional conditions are also favorable in the region beyond the delta, where at 2 m depth more than 25%, and at 3 m up to 39%, of the suspended and bed loads entering the delta are deposited (Table 5). Thus all but 16.3% of the Volga alluvium, mostly pelitic material, is found within the 3-m isobath.

In the relatively straightforward Don delta, according to Bronfman [2] as much as 12%, or 0.5 mill. tons, of sediment are deposited from a yearly supply to the delta of 3.94 and 0.2 mill. tons of suspended and bed loads, respectively. Deposition of sand and silt as the terrigenous material crosses the delta results in a progressive change in grain size (Table 6). High level of pelitic material in the river load are apparently associated with erosion of the floodplain.

Most of the suspended and bed load material is deposited on the shoreface in front of the delta. The easternmost part of the Gulf of Taganrog, 500 km², acts as a trap for sediments from the Don. According to Panov and Manykina [12], a layer of sediment 0.3-0.5 m thick, and locally as much as 1 m, formed here during 1903-1951 (average rate of deposition 3.3 mm/yr). Taking the bulk density of a most sample (1.54 g/cm³) and the moisture content (46%), and performing the appropriate calculations, we find that the amount of sediment deposited in the area of the mouth of the Don is 1.78 mill. tons/yr, which is about 47% of the river load (Table 7).

TABLE 9. Deposition of Various Size Fractions on the Delta and Shoreface of the Kuban, Mill. Tons:

Fraction, mm*	Supply of suspended and bed load during natur- al flow (1936- 1972, town of Tikhovskii)†	Deposition of terrigenous material		Carried to the Sea of Azov
		on delta	on shoreface	
< 0,01	3,95	1,23	0,71 ‡	2,01
0,01-0,05	1,68	0,50	0,59 **	0,59
0,05-0,25	0,91	0,22	0,69 ‡	-
> 0,25	0,19	0,08	0,11 ‡	-
Total	6,73	2,03	2,10	2,60

*Grain size data from [9, 15].

†Continuing bed load 20% of suspended load of the Kuban.

‡According to Simonov [15], about 15% of the suspended material accumulates on the shoreface, including all of the sand and coarse silt.

**Arbitrary assumed that half of the fine silt is deposited on the shoreface and half enters the sea.

The depositional environment in the delta and river mouth area of the Kuban is quite different. The many estuaries, steep submarine slope, the shoreline open to the sea, and the predominance of pelitic material in the river sediment affect the nature of the depositional regime. Unlike the Don, here there is active sedimentation on both the delta and shoreface (Table 8). Only about 39% of the load reaching the head of the delta, mainly the pelitic and fine silt fractions, is carried out to sea (Table 9).

Analysis of sedimentation in the foredelta areas of nontidal seas shows that the principal hydrodynamic parameters determining the disposition of the terrigenous material are current, waves, and movement of water onshore and offshore (Fig. 3). Depending on the predominant type of energy involved in the depositional environment, the foredelta area is characterized by a certain set of facies and bottom sediment lithotypes. In particular, sand, coarse silt, and shells are most common in those deposits where wave and onshore-offshore water movements are predominant (for example, the Kuban, Samur, Syrdar'ya, and small rivers). Areas where the solid load has an effect on the composition of bottom sediments vary, depending on their size. Often the shore is composed almost entirely of shelly material (for example, along the shoreface of the Kuban), derived from the floor of the Sea of Azov [10]. Shorefaces near rivers with a large discharge and solid load are different. In these cases, even clayey muds are formed, and most of the river's terrigenous material is deposited on the foredelta. The degree of influence of one or the other of the energy parameters is closely related to the configuration of the river mouth and delta, which varies considerably (Fig. 4).

The shape of the delta and processes of sedimentation depend to a considerable degree on the tectonic conditions. As shown by Richter [13], the most active delta formation is on uplifted shores. In such cases, the main tendency is for a rapid advance of the shoreline with minor deposition. The Volga delta is a classical example. Over a 10-yr period (1930-1940), the shoreline advanced seaward some 5-13 km [4], which, of course, was aided by a sharp fall in the level of the Caspian Sea. On the other hand, rapid accumulation of clastic material in Neocaspian time has proven insignificant (0.4-0.5 mm/yr) for the deltas studied.

Deltas on subsiding coastlines are characterized by a more complex pattern of development, which depends on the relationship between the rate of subsidence and the rate of deposition [13]. Studies have shown that high rates of sedimentation, reaching as much as 2-3 cm/yr, are typical. Significant differences have also been noted in the lithologic composition of the deltaic deposits formed under various tectonic conditions.

Depending on the ratio between the total amount of alluvial and marine sediment deposited along the shore, i.e., the rate of deposition, and the rate of epeirogenic subsidence, deltas in such regions can be subdivided according to three types of downwarping: undercompensated, compensated, and overcompensated. In undercompensated subsidence (for example, the Kuban delta), the principal effect is shoreline erosion, reaching as much as 5-6 m/yr, and minor accumulation of alluvium is found only in front of the main channel [25]. In its present stage of development, the Don delta is in a state of relative stability. Minor growth of the shoreline (averaging up to 6 m/yr) is maintained only in protected areas.

The deltas of the Amudar'ya and Syrdar'ya are examples of overcompensation. In these cases, subsidence of the sea floor is fully compensated by deposition of clastic material supplied by the rivers, which leads to active growth of the deltas. According to Zalogin and Rodionov [4], the seaward margin of the Syrdar'ya delta advances during natural flow an average of 100 m/yr. Thus, the main features of sedimentation and delta formation on uplifted coastlines are active advancement of the shoreline and a relatively low rate of accumulation of clastic material.

* * *

The foregoing discussion suggests that extremely active accumulation and reworking of sediments derived from land take place in the area of river mouths. Not by chance did Lisitsyn [7] refer river mouth areas to the primary global zone of slide deposits which should be considered large traps in which up to 80% of the clastic material is deposited. It is here that the greatest amount of mechanical differentiation takes place, separating the fine particles from the coarse. It differs in this respect from the river system, which is the area where the particles are derived and transported [19]. At the margin of the foredelta, where river and marine waters generally merge, a rapid fall-off in the river current together with an extremely high rate of deposition produce a poorly sorted, mainly terrigenous deposit, with a wide range of grain sizes from sand to fine silt and even clay, that is similar to alluvium in composition. The amount of sedimentation of the suspended and bed loads on the delta and adjacent shoreface is quite uneven, and is determined by the distribution of solid and liquid discharge in the channel, the nature of the submarine slope, and other morphological features as well as the hydrodynamic regime [26]. Moreover, the delta region can serve as an example of the lack of correspondence between hydrodynamic activity (kinetic energy) and the volume of sediment supplied by the river. It is entirely natural that the sediments here are typically low in sorting and varied on lithologic composition.

LITERATURE CITED

1. L. A. Barsukova, "Biogenic load of the Volga River before and after regulation of its flow (1936-1962)," *Annotatsia KaspNIIRKh*, vyp. 6, 91-131 (1965).
2. A. M. Bronfman, "Formation and anthropogenic alterations in the suspended load in the Sea of Azov," in: *Geographic Aspects of the Study of the Hydrology and Hydrochemistry of the Azov Basin* [in Russian], Izd. Geogr. o-va SSSR, Leningrad (1981), pp. 43-57.
3. A. M. Bronfman and A. N. Aleksandrov, "Sedimentation in marine shallows in front of a river mouth, in the case of the Don River," *Okeanologiya*, 5, vyp. 4, pp. 662-672 (1965).
4. B. S. Zalogin and N. A. Rodionov, *River-Mouth Regions of Soviet Rivers* [in Russian], Mysl' Moscow (1969).
5. K. I. Ivanov, "Deposition of suspended sediment on the shoreface near the mouth of the Kura River," *Tr. GOIN*, vyp. 28(40), 131-136 (1955).
6. J. M. Coleman and L. D. Wright, "Modern river deltas: variations in processes and sand bodies, in: *Deltas - Models for Study* [Russian translation], Nedra, Moscow (1979), pp. 32-91.
7. A. P. Lisitsyn, "Estuarine sedimentation," in: *Estuarine Sedimentation in the Ocean* [in Russian], Izd-vo Rostov. Univ., Rostov-on-Don (1982), pp. 3-58.
8. G. V. Lopatin, "Solid discharge of rivers in the Caspian Basin," *Tr. GGI*, vyp. 4(48), 78-95 (1948).
9. G. V. Lopatin, *River Sediments of the USSR (Origin and Transport)* [in Russian], Geografiz, Moscow (1952).

10. V. A. Mamykina and Yu. P. Khrustalev, Coastal Zone of the Sea of Azov [in Russian], Izd-vo Rostov. Univ., Rostov-on-Don (1980).
11. V. N. Mikhailov, "Dynamics of channel and current for non-tidal river mouths," Tr. GOIN, vyp. 102 (1971).
12. D. G. Panov and V. A. Mamykina, Rate of Deposition of Sediments in the Sea of Azov [in Russian], Avtoref. Nauch.-Issled. Rabot Rostov. Univ.; Izd-vo Rostov. Univ., Rostov-on-Don (1960).
13. V. G. Richter, Methods of Study of recent and Modern Tectonics on the Shelf Zone of Seas and Oceans [Russian translation], Nedra, Moscow (1965).
14. M. M. Rogov, S. S. Khodkin, and S. K. Revina, Hydrology of the Mouth of the Amudar'ya [in Russian], Gidrometeoizdat, Moscow (1968).
15. A. I. Simonov, Hydrology of the Mouth of the Kuban [in Russian], Gidrometeoizdat, Moscow (1958).
16. A. I. Simonov, Hydrology and Hydrochemistry of the River-Mouth Shoreface [in Russian], Gidrometeoizdat, Moscow (1969).
17. B. A. Skopintsev, "Coagulation of terrigenous material suspended in river discharge into seawater," Izd. Akad. Nauk SSSR, Ser. Geogr. Geofiz., 10, No. 4, pp. 357-371 (1946).
18. N. A. Skriptunov, Hydrology of the Shoreface near the Mouth of the Volga [in Russian], Gidrometeoizdat, Moscow (1958).
19. N. M. Strakhov, Basic Theory of Lithogenesis [in Russian], Izd-vo Akad. Nauk SSSR, Vols. 1-3, Moscow (1962).
20. N. M. Strakhov, N. G. Brodskaya, L. M. Knyazeva, et al., Formation of Sediments in Modern Basins [in Russian], Izd-vo Akad. Nauk SSSR, Moscow (1954).
21. Yu. P. Khrustalev, Patterns of Modern Deposition in the Northern Caspian [in Russian], Izd-vo Rostov. Univ., Rostov-on-Don (1978).
22. Yu. P. Khrustalev, Sedimentation in Areas Affected by River Currents [in Russian], Izd-vo Rostov. Univ., Rostov-on-Don (1978), pp. 59-71.
23. Yu. P. Khrustalev, Sedimentation in Epicontinental Seas in Arid Zones [in Russian], Doctoral Dissert., Geol.-Min. Sci., Inst. Okean., Akad. Nauk SSSR, Moscow (1986).
24. Yu. P. Khrustalev, S. A. Reznikov, and D. S. Turovskii, Lithology and Geochemistry of Bottom Sediments in the Aral' Sea [in Russian], Izd-vo Rostov. Univ., Rostov-on-Don (1977).
25. Yu. P. Khrustalev and F. A. Shcherbakov, Late Quaternary Deposits in the Sea of Azov and Their Environment of Deposition [in Russian], Izd-vo Rostov. Univ., Rostov-on-Don (1974).
26. A. A. Chistyakov, Environment of Deposition and Facies of Deltas and Deepwater Debris Fans [in Russian], Izd-vo VINITI, Moscow (1980).
27. G. I. Shamov, "Grain-size distribution of river alluvium in the USSR," Tr. GOIN, vyp. 18(72) (1951).
28. V. A. Sharapov, "Effect of regulating the river flow by reservoirs on nature and the economy downstream," Vopr. Geograf., No. 73, 136-152 (1968).
29. E. F. Shul'gina, "Distribution of certain hydrochemical and hydrological elements in the shallows off the Kuban estuary from observations made during August-September, 1951," Tr. GOIN, No. 28(40), 152-162 (1955).
30. V. L. Shul'ts, Rivers of Central Asia [in Russian], Gidrometeoizdat, Leningrad (1965).
31. K. R. Dyer, Sedimentation in Estuaries, in: the Marine Environment, R.S.K. Barnes and J. Z. Green (eds.), Appl. Sci. Publ., London (1972), pp. 10-32.
32. J. K. Edzwald, J. B. Upchurch, and C. R. O'Melia, "Coagulation in estuaries," Environm. Sci. Technol., 8, No. 1, 58-63 (1974).
33. H. H. Hahn and W. Stumm, "The role of coagulation in natural waters," Am. J. Sci., 268, 354-368 (1970).

In the present article, we discuss the principal factors responsible for the thermoluminescence of calcite and carbonate rocks. The effects of clay and anhydrite admixtures upon the γ -thermoluminescence (GTL) of carbonates are evaluated on the basis of experimental data. A scheme is proposed for delineating packets of GTL within a space containing different stratigraphic subdivisions. It is shown that the GTL method substantially enhances traditional biostratigraphic, lithological, and commercial-geophysical investigations of carbonate deposits. It allows sections to be analyzed and traced over areas of synchronous rock packets within carbonate deposits.

During geological exploration for oil and gas, the data from commercial-geophysical investigations (CGIs) serve as the main correlation for deposits and geological structures guiding exploratory efforts over the area. Correlation of carbonate deposits on the basis of CGI data alone, however, is often difficult and can result in substantial errors if one is comparing formations of complex lithofacial composition which contain no clear key horizons of proven uniform age and ubiquitous occurrence over the area. In the absence of characteristic fossil complexes (a situation often found in carbonate strata), the paleontological method is also insufficiently effective for analyzing and correlating such deposits.

Gamma-thermoluminescence (GTL) has been successfully used as a method to significantly broaden possibilities for analyzing carbonate deposits and tracing synchronous surfaces within thick, uniform limestone beds. The method is based upon the previously discovered [2] thermoluminescence of calcite when it is brought to a temperature below red heat (up to 400°C).

The thermoluminescence (TL) of calcite, responsible for the thermoluminescent properties of carbonate rocks is due to the presence of defects in the crystal lattice of the mineral formed under the influence of natural radiation and due to the incorporation of microelements consisting of ions which are activators or quenchers of TL into the lattice [5].

Terrestrial and cosmic radiation are the sources of the ionizing radiation. Cosmic radiation arises as a result of the bombardment of the upper layers of the atmosphere with elementary particles, primarily neutrons. In the process, nuclear reactions take place accompanied by secondary γ -radiation and the formation of significant quantities of radioactive elements. With increasing depth of burial, the level of ionizing radiation acting upon sediments decreases substantially as a result of absorption in the upper layers of the earth's surface.

Terrestrial radiation is created by the emissions of the radioactive elements involved in a complex chain of radioactive transformations with uranium-238 and thorium-232 as ancestors and potassium-40 and carbon-14 among the other important radioactive elements. The quantities of these radioactive elements within carbonates are very small under ordinary conditions (uranium-238 and thorium-232 constitute $10^{-4}\%$ and potassium-40 constitutes $\sim 1\%$). Moreover, the constants characterizing the decay rates of these elements are very small. For potassium-40, the constant equals $5.5 \cdot 10^{-10} \text{ g}^{-1}$. It is $1.5 \cdot 10^{-10} \text{ g}^{-1}$ for uranium-238 and $5.0 \cdot 10^{-11} \text{ g}^{-1}$ for thorium-232.

Taking into account the penetrative ability of ionizing radiation and the content of radioactive elements, it can be assumed that the probability of radiational defects arising after burial in addition to those created during the formation of the minerals is negligibly small.

Special investigations of biogenic carbonate rocks [1] showed an absence of stable relationships between the quantities of admixture elements accumulated in calcite and changes in the biotic environment of the skeletal lime-secreting organisms.

All-Union Scientific-Research Institute for Oil Exploration, Moscow. Translated from *Litologiya i Poleznye Iskopaemye*, No. 1, pp. 105-114, January-February, 1990. Original article submitted September 14, 1988.

Secondary dolomites do not show thermoluminescence. Dolomitized limestones are characterized by thermoluminescence from relict calcite.

The law that genetic information must be present within a mineral, since complete loss of such information would result in the mineral's destruction, insures the preservation of the primary thermoluminescence characteristics within the mineral during all of the stages in its postsedimentational transformation [8].

Carbonate rocks are heated when they are buried within the depths of the earth. This results in a "de-excitation" of the limestones, expressed in the disappearance of low-temperature (to 200°C) thermoluminescence. To restore the initial state of the crystalline lattice, powders prepared from samples of carbonate rocks can be subjected to ionization using γ -rays in doses of 150,000 roentgens, a dose sufficient for a second occupation of freed-up ("de-excited") microdefect-traps by electrons.

The γ -thermoluminescence analysis was carried out on a "Thermolum" (UITL-1) instrument. To obtain uniform results, the thickness of the layer of powder in the sample for study should equal the height of the skirting of the heat plate and the powder should be covered by a metal mask with a fixed opening providing an area of surface glow identical for all samples. Uniform heating of the samples to 400°C was continued for two minutes. This interval of time allowed us to obtain differentiated GTL-curves and to conduct a sufficiently massive treatment of the sample.

The γ -thermoluminescence of the sample, converted into electrical impulses, is recorded on a moving tape in the form of a continuous diagram. Flashes are recorded in the form of peaks of various size depending upon the intensity of the glow. The most characteristic GTL peaks are located in the following temperature intervals (Fig. 1): first peak) 80-100°C (low-temperature); second peak) 130-140°C; third peak) 170-180°C (intermediate-temperature); fourth peak) 230-240°C; fifth peak) 270-280°C; sixth peak) 310-320°C (high-temperature). If a diagram has three peaks of thermoluminescence, the second, fourth, and sixth peaks are the peaks most often absent. When the selected parameters of investigation are adhered to, GTL allows complete reproducibility for the results of control readings.

The authors investigated the effect of clay and anhydrite admixtures in limestones upon GTL characteristics. As the experimental investigations using artificially prepared powder mixtures of carbonate with different contents of clay impurity demonstrated, there is a distinct decrease in the intensity of the GTL with total preservation of the shapes of the luminescence peaks of the original sample on the diagrams.

The effect of anhydrite admixtures upon the form of the γ -thermoluminescence diagrams of limestones was investigated in powder mixtures of limestones containing admixtures in various proportions. As a result of the investigations, it was found that an admixture of anhydrite has practically no effect upon the shape of the GTL curve. Even for anhydrite in a mixture with limestone at a 5:1 ratio, the GTL diagram maintains the shape of a diagram for limestone (Fig. 2).

Various investigators have analyzed carbonate strata of the Devonian, Carboniferous, and Jurassic with the aid of the GTL method [3, 4, 6, 7]. These reports do not go beyond the framework of individual areas and, in each case, the indexing of the GTL packets was carried out without taking into consideration the possibilities of comparing the areas with sections from bordering regions. The authors developed a scale which allows one to correlate packets of rock identified on the basis of GTL data with a concrete, regional stratigraphic scale and to compare these scales with the general stratigraphic scale (Fig. 3). It should be noted that the delineation of GTL packets is carried out from the top downward, since the breakdown of the section is derived from well drilling and the wells often penetrate deposits of different ages even at small spacings.

At the present stage of stratigraphic study of the oil-and-gas territories within the USSR, the ages of deposits are known to within at least the range of a series, so several packets can be initially differentiated within a series (Table 1, a). In cases when deposits have stage subdivisions, packets of GTL make it possible to analyze deposits within a stage (see Table, b) and substage (see Table 1, c). A combined interpretation of the existing data from GTL and paleontology, and CGI make it possible to conduct, in a number of cases, fairly detailed analysis, up to the level of distinguishing packets of GTL within a horizon (see Table 1, d). Since the indexing encompasses all the known stratigraphic nomenclature, correlations can be made from more well known to less well investigated regions. The technique

being proposed makes it possible to establish within which age interval the more detailed analysis according to the GTL method should proceed and to compare packets within intervals of corresponding stratigraphic ranges, regardless of the local indexing.

The efficacy of the proposed method can be demonstrated in the example of investigations carried out on Carboniferous–Early Permian carbonate deposits of various facial types which were penetrated by deep boreholes within the Karachaganak reef complex in the northern portion of the pre-Caspian Basin.

The carbonate deposits in this area were subjected to a combination of paleontological, lithofacial, and γ -thermoluminescence investigations. Detailed paleontological investigations included study of the foraminifers, radiolarians, corals, brachiopods, conodonts, algae, spores, and pollen. The breakdown of the deposits into stages and horizons was made on the basis of the identified characteristic complexes of organic remains. The lithofacial investigations revealed the lithotypes present and the polyfacial nature of the deposits. They also allowed the facies variation to be traced laterally.

The stratigraphic units delineated over the area of Karachaganak correspond to analogous regional subdivisions of stratigraphic schemes employing the same names for the Carboniferous and Permian deposits of the Russian Platform proclaimed by the interdepartmental Stratigraphic Committee in 1988.

GTL studies of more than 2,000 samples along three parametric wells and 18 exploratory boreholes were carried out for the subsaline carbonate sediments of the Karachaganak deposit. Each interval characterized by core was studied. Selection of samples was carried out in stratigraphic succession, usually at intervals of 1 m, more rarely at intervals of 0.5 m. All of the lithological varieties of rock were samples. The majority of the samples had reliable stratigraphic control.

Auxiliary units (GTL packets) divided up to allow clarification of peculiarities in the internal structure of large stratigraphic taxons on the basis of changes in the thicknesses of packets of uniform age were delineated within the stratigraphic subdivisions using γ -thermoluminescence studies and taking into account paleontological and commercial geophysical data.

The Carboniferous deposits were divided into the Tournasian, Visean, Serpukhovian (lower division) and Baskhirian (upper division) stages, which were represented to an incomplete extent. The genetic types of the formations mentioned are especially diverse and include sediments of the shallow water shelf, submarine bank, reef complexes (bioherm, train, intra-

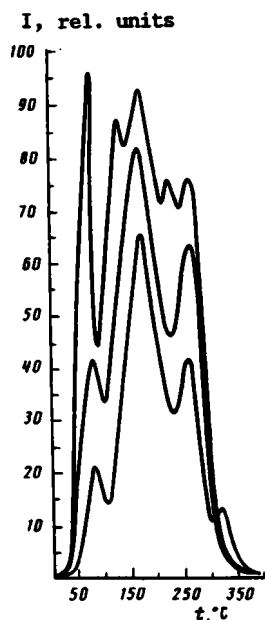


Fig. 1. Diagrams of the γ -thermoluminescence of carbonate rocks (I – intensity of GTL, relative units).

TABLE 1. Delineation of GTL Packets within the Space of the Various Stratigraphic Subdivisions

Devonian D	Carboniferous C	System
Upper D ₃	Lower C ₁	Series
D ₃ II D ₃ I	C ₁ II C ₁ I	Packet from GTL
Devonian D	Carboniferous C	System
Upper D ₃	Lower C ₁	Series
Famennian D ₃ fm	Tournaisian C ₁ t	Stage
Upper D ₃ fm ₃	Lower C ₁ t ₁	Substage
D ₃ fm ₃ 3 D ₃ fm ₂ 2 D ₃ fm ₃ 1	C ₁ t ₁ 3 C ₁ t ₁ 2 C ₁ t ₁ 1	Packet from GTL
Devonian D	Carboniferous C	System
Upper D ₃	Lower C ₁	Series
Famennian D ₃ m	Tournaisian C ₁ t	Stage
Upper D ₃ fm ₃	Lower C ₁ t ₁	Substage
D ₃ fm ₃ 3 D ₃ fm ₂ 2 D ₃ fm ₃ 1	C ₁ t ₁ 3 C ₁ t ₁ 2 C ₁ t ₁ 1	Packet from GTL
Devonian D	Carboniferous C	System
Upper D ₃	Lower C ₁	Series
Famennian D ₃ fm ₃	Tournaisian C ₁ t	Stage
Upper D ₃ fm ₃	Lower C ₁ t ₁	Substage
Ziganskii D ₃ fm ₃ zn	Gumerovskii C ₁ t ₁ gm	Horizon
D ₃ fm ₃ zn _c D ₃ fm ₃ zn _b D ₃ fm ₃ zn _a	C ₁ t ₁ gm _c C ₁ t ₁ gm _b C ₁ t ₁ gm _a	Packet from GTL

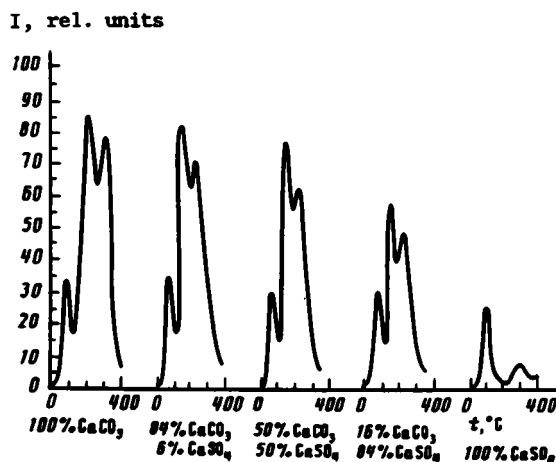


Fig. 2. GTL-diagrams of mixtures of limestone with anhydrite.

reef lagoon, and bar sediments), submarine slope, and deep waters (basinal sediments). Eight GTL packets of uniform age were found within the section of Carboniferous deposits and their distributions were traced over the area. Diagrams of GTL typical of each packet were considered time lines.

Two GTL packets can be distinguished going from bottom to top within the Tournaisian (represented by the lower substage and including the undivided Gumerovskii and Malevskii horizons); these are $C_{1t_1}^{gm+mlb}$ and $C_{1t_1}^{gm+m1a}$. The lithofacial compositions of the deposits are similar and indistinguishable on the basis of CGL data. The lithotypes in the packets of $C_{1t_1}^{gm+mlb}$ are represented by algal-spheroidal, spheroidal-ornamental, algal, organogenic-detrital, and detrital limestones.

GTL diagrams of $C_{1t_1}^{gm+mlb}$ packets are characterized by four peaks. The moderate-temperature second and third peaks are maximal in intensity. The low-temperature first peak is very weakly expressed.

The lithotypes in the $C_{1t_1}^{gm+m1a}$ packets include algal-spheroidal, spheroidal-ornamental, algal, organogenic-detrital, and coagulated limestones. GTL diagrams of the packet are distinguished by three peaks and a high intensity of thermoluminescence in the interval of intermediate temperatures. The low-temperature first peak is significantly smaller than the moderate-temperature third peak. The high temperature fifth peak has a minimum value.

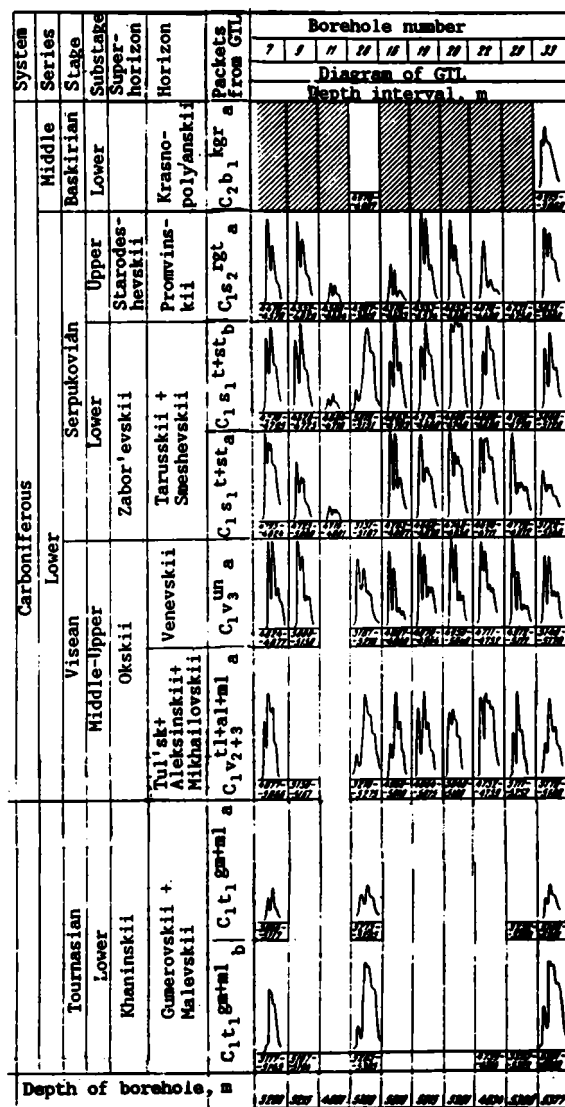


Fig. 3. Breakdown of Lower and Middle Carboniferous carbonate sediments in the Karachaganak deposit on the basis of GTL data.

The Visean Stage is limited to the upper portion of the middle and upper substages (Tul'skii, Aleksii, and Mikhailovskii horizons, and the Okskii super-horizon, and is divided into two types of GTL packets (from bottom to top): 1) $C_{1v_2-3}^{tl+al+mh}$ Tul'sk, Aleksii, and Mikhailovskii horizons, and 2) $C_{1v_3}^{vn}$, corresponding to the Venevskii horizon in most portions of the borehole sections, the undivided Okskii horizon can be distinguished on the basis of paleontological and lithofacies data, it is only possible to make distinctions at the horizon levels in certain sections. The lithofacies composition of the deposits mentioned is varied.

The lithotypes of the Okskii super-horizon are represented by organogenic, algal, foraminiferous, crinoidal, detrital, organogenic-detrital, oncoidal, and organogenic-fragmental limestones, calcareous claystones, aleuropelites, and siltstones.

A series of four-peak diagrams are characteristics of two of the delineated GTL packets (see Fig. 3). A minimal intensity of a fifth, high-temperature peak and a maximal intensity of the third, moderate-temperature peak are characteristic of the series of diagrams for the $C_{1v_2-3}^{tl+al+mh}$ packet. The intensities of the low-temperature first peak and the moderate-temperature fourth peak vary over similar ranges.

The series of diagrams for the $C_{1v_3}^{vn}$ packet is distinguished by maximal intensities of the low-temperature first and the moderate-temperature third peaks and minimal values for the high-temperature fourth and fifth peaks.

The Serpukhovian is represented on the basis of biostratographic data from the lower (the undivided Tarusskii and Steshevskii horizons) and the upper (Protvinskii horizon) sub-stages and can be divided in more detail into three packets on the basis of the γ -thermoluminescence analysis (bottom to top): 1) C_{1s}^{st+stb} ; 2) C_{1s}^{st+sta} , and 3) C_{1s}^{spgta} . The first and second packets correspond to the Tarusskii and Steshevskii horizons; the third packet corresponds to the Protvinskii horizon.

The lithotypes of the lower Serpukovian substage are represented by organogenic, organogenic-detrital, algal, crinoidal, foraminiferal, coral, detrital, oncoidal, oolitic, and microbrecciated limestones, secondary dolomites, and dolomitized limestones.

The series of diagrams for C_{1s}^{st+stb} are characterized by four peaks. The intensities of the moderate-temperature third and the high-temperature fourth peaks fluctuate from minimal values close to the value for the fifth, high-temperature peak to a maximum shown by the first, low-temperature peak.

The GTL packet C_{1s}^{st+sta} is distinguished by three-peak diagrams. The third moderate-temperature peak possesses maximum intensity. An analysis of the diagrams within this series indicates that they differ somewhat in intensity though they show an overall similarity. The decrease in the intensity of the high temperature fifth peak in the presence of a simultaneous increase in the intensity of the low-temperature first peak is probably due to the stratigraphic position of the samples studied, a position coinciding with the upper portion of the packet.

This conclusion is confirmed by the GTL-characterization of the overlying Protvinskii horizon.

The diagrams of the C_{1s}^{spgta} packets corresponding to the Protvinskii horizon are also three-peaked. The intensities of the peaks successively fall from the low-temperature first peak to the high temperature fifth peak.

The GTL-diagrams of the $C_{2b_1}^{kgra}$ packets corresponding to the Krasnopolyanskii horizon of the Visean are distinguished by a peculiar five-peaked shape. Maximum intensities of the thermoluminescence are characteristic of the two intermediate-temperature peaks. Minimum intensity is characteristic of the first, low-temperature peak.

A comparative analysis of the GTL characteristics of the Carboniferous deposits in the region under consideration shows that there is a transition within a stage (substage) from a complex, 4-peak form of diagram to a simpler, 3-peak configuration upward along the section in the Tournaisian, Visean, and Serpukhovian formations. Complex, slightly disrupted, intensive 5-peaked diagrams are characteristic of the lower portion of the Krasnopolyanskii horizon within the Baskhirian of the Middle Carboniferous.

The Lower Permian (pre-Kungurian) deposits overlapped by discontinuous Lower- and Middle-Carboniferous rocks can be divided into Asselian, Sakmarian, and Artinskian stages on the basis of biostratographic reports. The genetic types are represented by reef, lagoonal, slope, and relatively deep water (basinal) formations.

The lithotypes of the Asselian stage include organogenic, bryozoan-fusulinid, bryozoan, bryozoan-detrital, fusulinid, algal, tubiphytic, crinoidal-fusulinid, coral-bryozoan, crinoidal, organogenic-detrital, detrital, coagulated, finely and very finely crystallized limestones with relict-organogenic diagenetic dolomites, and calcareous siltstones and claystones.

Three packets can be distinguished in the Asselian stage on the basis of γ -thermoluminescence studies (bottom to top): 1) P_{1aC} ; 2) P_{1aB} ; 3) P_{1aA} (Fig. 4).

The GTL packet P_{1aC} is characterized by 4-peak diagrams. The intensity of γ -thermoluminescence diminishes from the third intermediate-temperature peak to the fifth high temperature peak. The size of the first low-temperature peak is not constant.

The diagrams of packet P_{1aB} are 3-peaked, with similar thermoluminescence intensities at moderate temperatures.

The diagrams of packet P_{1aA} are also 3-peaked, but the intensity of the low-temperature first peak is minimal, while the intensity of the intermediate-temperature third peak is maximal. The high temperature fifth peak is higher than the low-temperature first, but lower than the intermediate-temperature third peak.

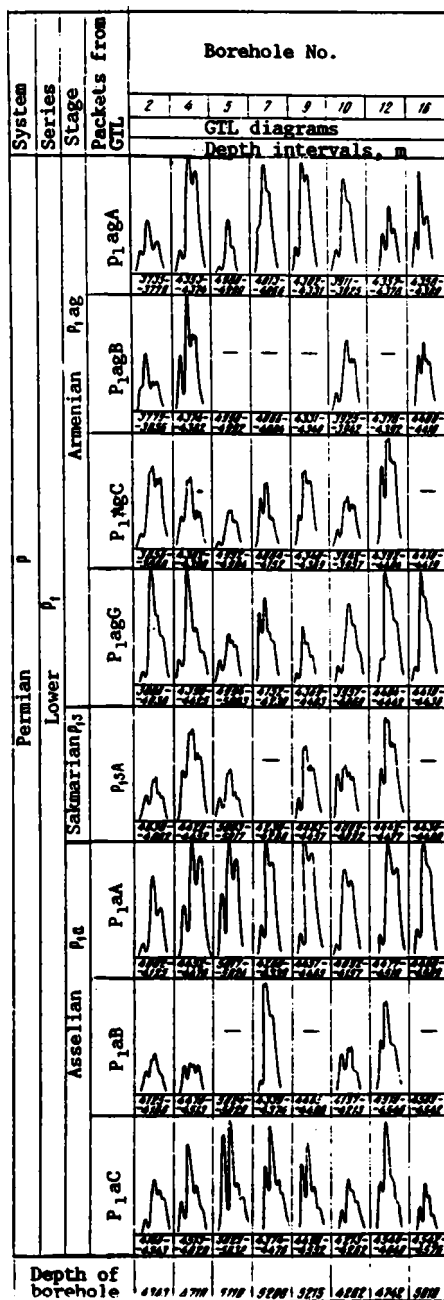


Fig. 4. Analysis of the Lower Permian carbonate sediments of the Karachaganak deposit of the basis of GTL data.

Consequently, complex 4-peaked diagrams are characteristic of the lower and middle portions of the Asselian stage, while morphologically simpler 3-peak diagrams are characteristic of the upper portion.

The Sakmarian stage combines the following lithotypes: organogenic, algal, bryozoan, tubiphytic, crinoidal-algal, organogenic-detrital, organogenic-fragmental, brecciated, finely crystalline, and pelitomorphous limestones, siltstones and aleuropelites. A GTL packet corresponds to this stage. The diagrams of the packet are complex in construction and 5-peaked. The thermoluminescence characterization of the limestones is distinguished by the presence of weakly expressed moderate- and low-temperature peaks which are not always clearly expressed in the presence of a clay admixture within the limestones.

The Artinskian stage is divided into four GTL packets (bottom to top). 1) P₁agG; 2) P₁agC; 3) P₁agB; 4) P₁agA (see Fig. 4). The lithotypes of this stage include organogenic,

algal, bryozoan, bryozoan-fusulinid, coral-algal, coral, organogenic-detrital, organogenic-fragmental, brecciated, and crinoidal-fusulinid limestones, spongiolites, relict-organogenic dolomites, and calcareous siltstones and aleuropelites.

Diagrams of the P_{1agG} packet are 4-peaked and are similar to the diagrams for the Asselian packet; they differ in having low thermoluminescence values for the low-temperature first peak.

Diagrams of the P_{1agC} packet are distinguished by five peaks. The intermediate-temperature peaks are similar in value and are separated by shallow troughs. An analogous pattern is noticeable for the high temperature peaks.

A series of 4-peak diagrams are characteristic of the P_{1agB} GTL packet. In comparison with the underlying packet, there is an absence of a second low-temperature peak in the diagrams.

Three peaks are characteristic of the series of diagrams for the P_{1agA} packet. The third intermediate-temperature peak is the maximum; the minimum is the first low-temperature peak. The total intensity of thermoluminescence varies significantly depending upon the quantity of clay admixture and anhydrite.

The comparison with the GTL diagrams of the pre-Kungurian limestones from the Lower Permian reveals certain peculiarities. The diagrams become more simple upward along this section: 4-peaked diagrams are replaced by 3-peaked diagrams in the Asselian stage and 5-peaked diagrams are replaced by 3-peaked diagrams in the Sakmariian-Artinskian deposits.

The first effort to analyze the carbonate section of the Karachaganak area, interpreted using significant quantities of core material, showed that each packet delineated on the basis of GTL data has individual characteristics and can be traced well over the area. The GTL characteristics of individual packets within the vertical sections of the deposits can be similar. The convergence is connected with the periodic repetition of environmental factors controlling the type of deformation of the crystalline lattice and, as a consequence, the form of the GTL diagram.

A detailed analysis and correlation of the borehole sections of the Karachaganak area based on GTL data in combination with paleontological, lithological, and commercial-geophysical investigations allowed us to create a geological model for the Karachaganak deposit.

When conducting the γ -thermoluminescence investigations, special attention was devoted to clarifying the degree of reliability of the GTL method compared with data from biostratigraphic reports. Comparison of the results obtained showed: 1) GTL data and the data of paleontology coinciding; changes in the GTL characteristics of auxiliary units distinguished at levels corresponding to the accepted boundaries of stratigraphic units; 2) auxiliary units established on the basis of GTL showing appreciably more detail than those distinguished on the basis of paleontological data; 3) the boundaries of units based on GTL data not coinciding with the generally accepted boundaries of units delineated on the basis of paleontological data.

The first and second situations are encountered the most frequently. An unequivocality of stratigraphic spaces occupied by primary (biostratigraphic) and auxiliary (GTL packets) units and a more detailed breakdown on the basis of GTL units usually raise no objections. The third situation is the most complex. Additional, more in-depth paleontological investigations are usually necessary to determine the reliability of the units delineated. As a rule, the results of such investigations further confirm the positions of boundaries based on GTL data and allow us to refine the stratigraphic schemes involved.

The γ -thermoluminescence analysis of limestones substantially enhances traditional biostratigraphic, lithofacies, and commercial-geophysical investigations of carbonate rocks.

The method allows one to resolve problems concerning the internal structure of strata and the structural relationships between complexes with deposits of diverse facies on the basis of a detailed analysis and areal tracing of synchronous surfaces within fixed strata of carbonate deposits insufficiently characterized by remains of guide fossils. Questions of interregional correlation of carbonate deposits, geological problems associated with finding prospective targets of the non-anticlinal type, and problems associated with paleostructural and other geological reconstructions can be solved on the basis of the method.

LITERATURE CITED

1. K. X. Wolf, J. V. Chilingar, and F. U. Biles, "Elemental composition of carbonate organic remains, minerals, and sediments," in: Carbonate Rocks [Russian translation], Mir, Moscow (1971), pp. 34-44.
2. Z. F. Daniel's, E. Bond, and D. Saunders, "Thermoluminescence as a means of scientific investigation," Usp. Fiz. Nauk, 51, 2, 271-286 (1953).
3. V. D. Il'in and V. N. Rumakin, "The use of the method of γ -thermoluminescence for correlating carbonate rocks of the Upper Jurassic in Western Uzbekistan," Tr. Vse-Soyuz. Neft. Nauchno-Issled. Geol. Inst., No. 95, 22-30 (1970).
4. M. V. Korovkin, V. N. Sal'nikov, A. F. Senakolis, et al., "Analysis of the carbonate strata of the Paleozoic basement of the oil-and-gas deposits of the Tomsk region using the method of γ -thermoluminescence," in: Problems in Geology and Exploration of Deposits of Commercial Resources of Siberia (Text of Reports at the Conference Celebrating the 100th Anniversary of the Birth of Academician M. A. Usov) [in Russian], In-t Geol. Geofiz. Siber. Otdel. Akad. Nauk SSSR (1983), pp. 15-18.
5. U. Medlin, "The nature of traps and emission centers in thermoluminescent rocks," in: The Physics of Minerals [Russian translation], Mir, Moscow (1971), pp. 55-59.
6. V. I. Tyurin and L. V. Sharonov, "An attempt to apply the method of γ -thermoluminescence to the correlation of the carbonate deposits of the Upper Devonian and Tournaisian stage of the Carboniferous in the Permsk region," in: The Stratigraphy, Lithology, Facies, and Fauna of the Upper Proterozoic and Paleozoic of the Volgo-Ural Oil-and-Gas Region [in Russian], Nedra, Moscow (1966), pp. 128-136.
7. G. E. Shinkarev and E. D. Abramova, "Feature associated with the thermoluminescence properties of the Famennian terrigenous-carbonate deposits of Central Karatau," in: Typomorphic Features and Technological Properties of the Minerals of Ore and Non-Ore Deposits of Kazakhstan and Their Practical Value [in Russian], Alma-Ata, Nauka (1984), pp. 88-91.
8. N. P. Yushkin, Theory and Methods of Mineralogy (Selected Problems) [in Russian], Nauka, Leningrad (1977).

CHANGING YOUR ADDRESS?

In order to receive your journal without interruption, please complete this change of address notice and forward to the Publisher, 60 days in advance, if possible.

(Please Print)

Old Address:

name

address

city

state (or country)

zip code

New Address

name

address

city

state (or country)

zip code

date new address effective

name of journal



233 Spring Street, New York, New York 10013

MECHANICS OF COMPOSITE MATERIALS*Mekhanika Kompozitnykh Materialov*

Vol. 26, 1990 (6 issues) \$745

MELTS*Rasplavy*

Vol. 4, 1990 (6 issues) \$295

METAL SCIENCE AND HEAT TREATMENT*Metallovedenie i Termicheskaya Obrabotka Metallov*

Vol. 32, 1990 (12 issues) \$915

METALLURGIST*Metallurg*

Vol. 34, 1990 (12 issues) \$915

PROBLEMS OF INFORMATION TRANSMISSION*Problemy Peredachi Informatsii*

Vol. 26, 1990 (4 issues) \$695

PROGRAMMING AND COMPUTER SOFTWARE*Programmirovaniye*

Vol. 16, 1990 (6 issues) \$395

PROTECTION OF METALS*Zashchita Metallov*

Vol. 26, 1990 (6 issues) \$915

RADIOPHYSICS AND QUANTUM ELECTRONICS*Izvestiya Vysshikh Uchebnykh Zavedenii, Radiofizika*

Vol. 33, 1990 (12 issues) \$915

REFRACTORIES*Ogneupory*

Vol. 31, 1990 (12 issues) \$815

SIBERIAN MATHEMATICAL JOURNAL*Sibirskii Matematicheskii Zhurnal*

Vol. 31, 1990 (6 issues) \$1045

SOIL MECHANICS AND**FOUNDATION ENGINEERING***Osnovaniya, Fundamenty i Mekhanika Gruntov*

Vol. 27, 1990 (6 issues) \$815

SOLAR SYSTEM RESEARCH*Astronomicheskii Vestnik*

Vol. 24, 1990 (4 issues) \$615

SOVIET APPLIED MECHANICS*Prikladnaya Mekhanika*

Vol. 26, 1990 (12 issues) \$915

SOVIET ATOMIC ENERGY*Atomnaya Energiya*

Vols. 68-69, 1990 (12 issues) \$945

**SOVIET JOURNAL OF GLASS PHYSICS
AND CHEMISTRY***Fizika i Khimiya Stekla*

Vol. 16, 1990 (6 issues) \$515

**SOVIET JOURNAL OF
NONDESTRUCTIVE TESTING***Defektoskopiya*

Vol. 26, 1990 (12 issues) \$1045

SOVIET MATERIALS SCIENCE*Fiziko-khimicheskaya Mekhanika Materialov*

Vol. 26, 1990 (6 issues) \$815

SOVIET MICROELECTRONICS*Mikroelektronika*

Vol. 19, 1990 (6 issues) \$545

SOVIET MINING SCIENCE*Fiziko-tehnicheskie Problemy Razrabotki**Poleznykh Iskopaemykh*

Vol. 26, 1990 (6 issues) \$855

SOVIET PHYSICS JOURNAL*Izvestiya Vysshikh Uchebnykh Zavedenii, Fizika*

Vol. 33, 1990 (12 issues) \$915

**SOVIET POWDER METALLURGY AND
METAL CERAMICS***Poroshkovaya Metallurgiya*

Vol. 29, 1990 (12 issues) \$945

STRENGTH OF MATERIALS*Problemy Prochnosti*

Vol. 22, 1990 (12 issues) \$1045

THEORETICAL AND MATHEMATICAL PHYSICS*Teoreticheskaya i Matematicheskaya Fizika*

Vols. 82-85, 1990 (12 issues) \$895

UKRAINIAN MATHEMATICAL JOURNAL*Ukrainskii Matematicheskii Zhurnal*

Vol. 42, 1990 (12 issues) \$995

Send for Your Free Examination Copy**Plenum Publishing Corporation**, 233 Spring St., New York, N.Y. 10013
In United Kingdom: 88/90 Middlesex St., London E1 7EZ, England

Prices slightly higher outside the U.S. Prices subject to change without notice.

RUSSIAN JOURNALS IN THE PHYSICAL AND MATHEMATICAL SCIENCES

AVAILABLE IN ENGLISH TRANSLATION

ALGEBRA AND LOGIC

Algebra i Logika

Vol. 29, 1990 (6 issues) \$635

ASTROPHYSICS

Astrofizika

Vols. 32-33, 1990 (6 issues) \$745

AUTOMATION AND REMOTE CONTROL

Avtomatika i Telemekhanika

Vol. 51, 1990 (24 issues) \$1045

COMBUSTION, EXPLOSION, AND SHOCK WAVES

Fizika Goreniya i Vzryva

Vol. 26, 1990 (6 issues) \$815

COMPUTATIONAL MATHEMATICS AND MODELING

A translation of selected works in mathematics

Vol. 1, 1990 (4 issues) \$295

COSMIC RESEARCH

Kosmicheskie Issledovaniya

Vol. 28, 1990 (6 issues) \$915

CYBERNETICS

Kibernetika

Vol. 26, 1990 (6 issues) \$815

DIFFERENTIAL EQUATIONS

Differentsial'nye Uravneniya

Vol. 26, 1990 (12 issues) \$915

DOKLADY BIOPHYSICS

Doklady Akademii Nauk SSSR

Vols. 310-315, 1990 (2 issues) \$275

FLUID DYNAMICS

Izvestiya Akademii Nauk SSSR.

Mekhanika Zhidkosti i Gaza

Vol. 25, 1990 (6 issues) \$915

FUNCTIONAL ANALYSIS AND ITS APPLICATIONS

Funktsional'nyi Analiz i Ego Prilozheniya

Vol. 24, 1990 (4 issues) \$695

GLASS AND CERAMICS

Steklo i Keramika

Vol. 47, 1990 (12 issues) \$945

HIGH TEMPERATURE

Teplofizika Vysokikh Temperatur

Vol. 28, 1990 (6 issues) \$915

HYDROTECHNICAL CONSTRUCTION

Gidrotekhnicheskoe Stroitel'stvo

Vol. 24, 1990 (12 issues) \$695

INDUSTRIAL LABORATORY

Zavodskaya Laboratoriya

Vol. 56, 1990 (12 issues) \$915

INSTRUMENTS AND EXPERIMENTAL TECHNIQUES

Pribory i Tekhnika Eksperimenta

Vol. 33, 1990 (12 issues) \$1015

JOURNAL OF APPLIED MECHANICS AND TECHNICAL PHYSICS

Zhurnal Prikladnoi Mekhaniki i Tekhnicheskoi Fiziki

Vol. 31, 1990 (6 issues) \$915

JOURNAL OF APPLIED SPECTROSCOPY

Zhurnal Prikladnoi Spektroskopii

Vols. 52-53, 1990 (12 issues) \$935

JOURNAL OF ENGINEERING PHYSICS

Inzhenerno-fizicheskii Zhurnal

Vols. 58-59, 1990 (12 issues) \$935

JOURNAL OF SOVIET LASER RESEARCH

A translation of articles based on the best

Soviet research in the field of lasers

Vol. 11, 1990 (6 issues) \$395

JOURNAL OF SOVIET MATHEMATICS

*A translation of selected Russian works
in mathematics*

Vols. 48-52, 1990 (30 issues) \$2250

LITHOLOGY AND MINERAL RESOURCES

Litologiya i Poleznye Iskopaemye

Vol. 25, 1990 (6 issues) \$915

LITHUANIAN MATHEMATICAL JOURNAL

Litovskii Matematicheskii Sbornik

Vol. 30, 1990 (4 issues) \$545

MAGNETOHYDRODYNAMICS

Magnitnaya Gidrodinamika

Vol. 26, 1990 (4 issues) \$695

MATHEMATICAL NOTES

Matematicheskie Zametki

Vols. 47-48, 1990 (12 issues) \$915

MEASUREMENT TECHNIQUES

Izmeritel'naya Tekhnika

Vol. 33, 1990 (12 issues) \$915

continued on inside back cover